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THE
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DEEP-FOCUS EARTHQUAKES AND THEIR GEOLOGICAL SIGNIFICANCE

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Madison, Wisconsin

AND

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Los Angeles, California

ABSTRACT

About 10 per cent of all major earthquakes are now known to originate below the "crust" of the earth at depths ranging up to 700 km. After a brief résumé of the evidence for the existence of deep earthquakes, the paper describes the environment in which they occur under the headings of structure and composition, density and pressure, rigidity and bulk modulus, temperature, viscosity, and strength. The characteristics of deep earthquakes are then described under the headings of distribution (geographic and depth), energy released, aftershocks, periodicities, correlations, patterns of initial motion at the surface, and stress and strain near the focal region.

Analysis of these data leads first to the conclusion that the type of process involved is probably a normal fracturing response to external stress differences, although other possibilities are considered. The possible known sources of stress appear to be inadequate to function as primary causes of deep earthquakes but may be important as "triggers." The meaning of strength is analyzed, and the conclusion is reached that seismic evidence probably points to an increase in strength from the surface of the earth downward at least to the core.

The paper concludes that the concept of a weak substratum, within which flow results from extremely slight stress differences, does not appear to be valid.

INTRODUCTION

The geological history of the earth is one of constantly recurring movements, at times of great magnitude, at times scarcely noticeable, but never, apparently, absent. It is naturally accepted that surface movements are related in some manner to movements at greater depths, and thus it is that any theory of diastrophism is

founded on a primary postulate concerning the nature of the depths of the earth. Owing to the unfortunate limitation of the zone of direct geologic observation to depths of a few kilometers, so few facts are known regarding the nature of the material at greater depths that the theorist has had a wide range of choices for his fundamental assumption. The result has been that concepts have been advanced and supported vigorously, not because of known facts concerning the subcrustal material, but rather because the concepts were necessary to the subsequent elaboration of a theory of diastrophism.

Although the primary assumption may have been chosen on the basis of expediency, the resulting theory has frequently been carried to satisfactory explanations of the surface features of the earth. An outstanding example presents itself in the assumption of the existence of a mobile substratum. This assumption permits isostatic compensation of crustal segments, the accommodation of great crustal shortening by the folding, downwarping, and foundering of the crystalline crust, and continental migration. It should be emphasized, however, that, if the acquisition of new data concerning the properties of the material of the interior of the earth should disprove the basic assumption of a theory, the explanations which follow from it, however satisfactory they may appear, must be entirely fortuitous.

In recent years a new source of information concerning the properties of the material within the earth has arisen with the discovery of the existence of earthquakes which originate at great depths beneath the surface. The first part of this paper summarizes the prevalent opinions concerning the interior of the earth, which are based largely on the study of shallow earthquakes and earthquake waves, and presents recently acquired data on the characteristics of deep-focus earthquakes. The second part attempts to evaluate the geological significance of these data. Definite limits to the permissible range of speculative concepts about the material within the earth are shown to be established by the existence and characteristics of deep earthquakes. These new data, analyzed in conjunction with the factors affecting the strength and modes of yield of a material,

strongly suggest the necessity of abandoning the traditional concept of an asthenosphere.

The authors wish to acknowledge their indebtedness to the following, who have aided in the preparation of this paper by providing data or stimulating discussions of the problem: Professors A. N. Winchell, L. B. Slichter, C. K. Leith, W. J. Mead, Dr. H. T. Stetson, Commander N. H. Heck, Rev. J. B. Macelwane, S.J., Dr. C. F. Richter, and Professor B. Gutenberg.

GENERAL RESULTS OF SEISMOLOGY

TERMINOLOGY

An *earthquake* is the rapid release of energy in the form of elastic waves. The region in which the process causing the energy release occurs is the "focal region," and the center of this region is the "focus." The region on the surface of the earth immediately above the focal region is the "epicentral region," and the point immediately above the focus, the "epicenter." At the instant of occurrence of an earthquake, elastic waves are produced and propagated outwards from the focal region in all directions. The waves which are recorded by a seismograph at some distance from the epicenter are "body waves," which have traveled in the interior of the earth, and "surface waves," which have traveled from the neighborhood of the epicenter around the surface of the earth to the location of the seismograph. The body waves are of two sorts: "compressional waves," in which the motion of a particle traversed by the waves is in the line of propagation of the waves; and "shear" or "transverse" waves, in which the motion is at right angles to the line of propagation.

WAVE VELOCITIES

The velocity of elastic waves depends on the density and elastic properties of the material through which they are passing. The velocity of compressional waves depends on both the bulk modulus (incompressibility) and the rigidity in accordance with the formula $v=[(k+\frac{4}{3}m)/p]^{\frac{1}{2}}$, where v is the velocity, k is the bulk modulus, m is the rigidity, and p is the density. The velocity of transverse waves depends on only one elastic property, the rigidity, according to the formula, $v=(m/p)^{\frac{1}{2}}$. It is clear that increasing rigidity has

the effect of increasing velocity, and increasing density has the effect of decreasing velocity. The velocity of compressional waves is greater than that of transverse waves, and, in the earth, the ratio of the velocities of the two waves is always very nearly 1.8:1. The velocity of surface waves is slightly less than that of transverse waves near the surface. Owing to their different velocities, the three kinds of waves arrive at a seismograph at different times, thus providing the characteristic seismogram, consisting of a primary (*P*) phase, corresponding to the arrival of the compressional wave; a secondary (*S*) phase, corresponding to the arrival of the transverse wave; and a phase of long waves (*L*), corresponding to the arrival of the surface waves. This latter phase normally has by far the largest amplitude on the seismogram.

STRUCTURE OF THE EARTH

From studies of the waves resulting from earthquakes or large commercial blasts, a considerable amount of information has been obtained concerning the structure of the outer portion of the earth. Such investigations indicate that beneath the superficial sedimentary layers there are two or more shallow layers, in each of which the velocity of elastic waves is practically constant. The upper layer, commonly called the "granitic layer," may be thin, or even absent, under portions of the ocean basins. At the bottom of the lower layer there is a velocity discontinuity where the velocity of compressional waves suddenly increases from 6–7 km./sec. to about 8 km./sec. Determinations of the depth of this discontinuity, called the "Mohorovičić discontinuity," range from 30 to 60 km., with recent determinations lying in the lower end of this range.

The portion of the earth above the Mohorovičić discontinuity is frequently called the "crust." The use of this term, though in many respects unfortunately chosen, is justified so long as we do not allow it to suggest any properties of the material within or below it other than the ability to transmit elastic waves with certain velocities. The great majority of earthquakes have their origin within the crust.

Analysis of the travel-time curves of seismic waves gives their velocities at all depths down to about half the radius of the earth, and average velocities at greater depths. The velocity of compressional waves is found to increase from about 8 km./sec. at the base

of the crust to about 13.5 km./sec. at a depth of 2,700 km. At this depth a decrease begins, continuing to 2,900 km., where the velocity is about 12.5 km./sec. At this depth there is a sharp discontinuity marking the boundary of the so-called "core," within which the velocity averages only about 10 km./sec. It is still an open question as to whether the core possesses sufficient rigidity to be capable of transmitting transverse elastic waves.

The earth's "mantle," that zone lying below the crust and above the core, does not contain any major discontinuities comparable to the Mohorovičić discontinuity or the discontinuity at the core, but it may contain several minor discontinuities. There is undoubtedly a second-order discontinuity (a discontinuity in the rate of change of velocity not capable of producing wave reflection) at the depth of about 1,000 km. Whether or not there are other minor discontinuities between this depth and the crust has not been determined definitely, evidence existing for both sides of the question.[1]

DEEP-FOCUS EARTHQUAKES

Owing largely to the very natural feeling that the material of the earth as a result of increasing temperature should become increasingly plastic with depth, seismologists and geologists have believed until comparatively recently that all earthquake foci are confined to the crust. The scarcity of seismic observations indicating the contrary, and the demand of many of the interpreters of surface structures and of the observations of gravitational anomalies for an asthenosphere, or zone of weakness, within which elastic stresses could not be maintained, combined to make this belief very strong.

In 1922 Turner, in the course of his work of editing the *International Seismological Summary*, observed that the waves from some earthquakes arrived at distant stations much too early. He ascribed this phenomenon to an abnormal depth of focus and devised a method of computing the depth of focus from the arrival times of the waves at distant stations.[2] Some foci were calculated by Turner to have been located several hundred kilometers below normal depth.

[1] B. Gutenberg and C. F. Richter, *Seis. Soc. Amer. Bull. 21* (1931), pp. 216–22; F. Neumann, *Trans. Amer. Geophys. Union, 14th Meeting* (1933), pp. 295–335; P. Byerly, *Seis. Soc. Amer. Bull. 16* (1926); *ibid., Bull. 24* (1934), pp. 81–99; *ibid., Bull. 25* (1935), pp. 62–80.

[2] H. H. Turner, *Monthly Notices Royal Astronomical Society, Geophys. Suppl.*, Vol. I (1922), pp. 1–13.

Seismologists were reluctant to accept the idea that earthquakes originate at such tremendous depths until the publication of a series of studies by Wadati,[3] beginning in 1928. Wadati found near Japan a class of earthquakes characterized by numerous phenomena which could be accounted for only by the assumption that these earthquakes had abnormally deep foci. Among the observations that demanded this were the anomalously widespread areas over which the shaking was sensible (although never destructive), an anomalously high apparent surface velocity of the earthquake waves, and a large interval between the times of arrival of the *P* and *S* phases at the epicenter. From this interval the depth of focus could be computed.

Other investigators soon began to make an intensive study of these unusual earthquakes. Scrase[4] found phases in their seismograms which he identified as waves which had been reflected from the surface of the earth near the epicenter, and which he denoted by the symbols *pP*, *sP*, *pS*, and *sS*. The arrival times of these phases checked well with the times predicted from the focal depths which had been assigned by Turner. Scrase also noted that the surface waves were abnormally small in amplitude—a characteristic which had been predicted theoretically by Banerji.[5] Stechschulte[6] demonstrated the usefulness of Scrase's phases in the determination of the depth of the focus from the seismograms at a single station. Brunner[7] constructed a convenient chart by means of which the epicentral distance of an earthquake from a given station and the focal depth can be read rapidly and accurately from a seismogram.

Thus within the past few years it has been established as a fact that earthquakes occur at what had been considered prohibitively great depths beneath the surface of the earth, and several methods have been devised for the accurate location of the foci of such earthquakes. It now appears that at least 10 per cent of all earthquakes originate at abnormal depths, i.e., below the crust. To date, the deepest earthquake which has been observed was that of June 29,

[3] K. Wadati, *Geophys. Mag.*, Vol. I (1928), pp. 168–202; *ibid.*, Vol. II (1929), pp. 1–36; *ibid.*, Vol. IV (1931), pp. 231–83.

[4] F. J. Scrase, *Proc. Roy. Soc.*, Vol. CXXXII, Part A (1931), pp. 213–35.

[5] F. J. Banerji, *Phil. Mag.*, Vol. XLIX (1925), pp. 65–80.

[6] V. C. Stechschulte, *Seis. Soc. Amer. Bull. 22* (1932), pp. 81–137.

[7] G. J. Brunner, *Trans. Amer. Geophy. Union, 15th Meeting* (1934), p. 77.

1934, the epicenter of which was a short distance southeast of the island of Celebes. The focus of this earthquake was at the tremendous depth of 700 km.,[8] which is one-ninth the radius of the earth.

Figure 1 is presented to give a comparison between the sizes of irregularities on the surface of the earth, the thickness of the crust, and the depths at which earthquake foci are observed to occur.

THE ENVIRONMENT OF DEEP-FOCUS EARTHQUAKES

STRUCTURE AND COMPOSITION

We shall now consider the properties which have been ascribed to the material of the mantle of the earth, i.e., the environment, in a general sense, of the range of depth within which earthquakes have been observed to originate.

It has already been pointed out that the mantle of the earth, to a depth of 1,000 km., is characterized by a lack of definite structural discontinuities. The limits to the information concerning the interior of the earth that can be obtained from the study of the arrival times of seismic waves should be emphasized. Such investigation gives the variation of velocity with depth but cannot separate the elastic constants and the density or tell anything directly about the composition of the material of the earth, except at very shallow depths in the crust. However, certain results can be inferred from the velocity formulas. The observed increase of velocity with depth indicates that the elastic constants (rigidity and bulk modulus) are increasing at a slightly greater rate than the density. Considering the slow, apparently continuous, variation of velocity with depth, it appears safe to assume that the composition of the mantle to a depth of 1,000 km. likewise undergoes only gradual changes.

DENSITY AND PRESSURE

An infinite number of modes of distribution of the density of the material of the earth with depth, each consistent with the known mass and moment of inertia of the earth, is possible. The choice is somewhat limited by taking into account the effect of the core on

[8] G. J. Brunner, *Prelim. Bull. 25a, Central Station of the Jesuit Seis. Assoc.* (St. Louis, 1934).

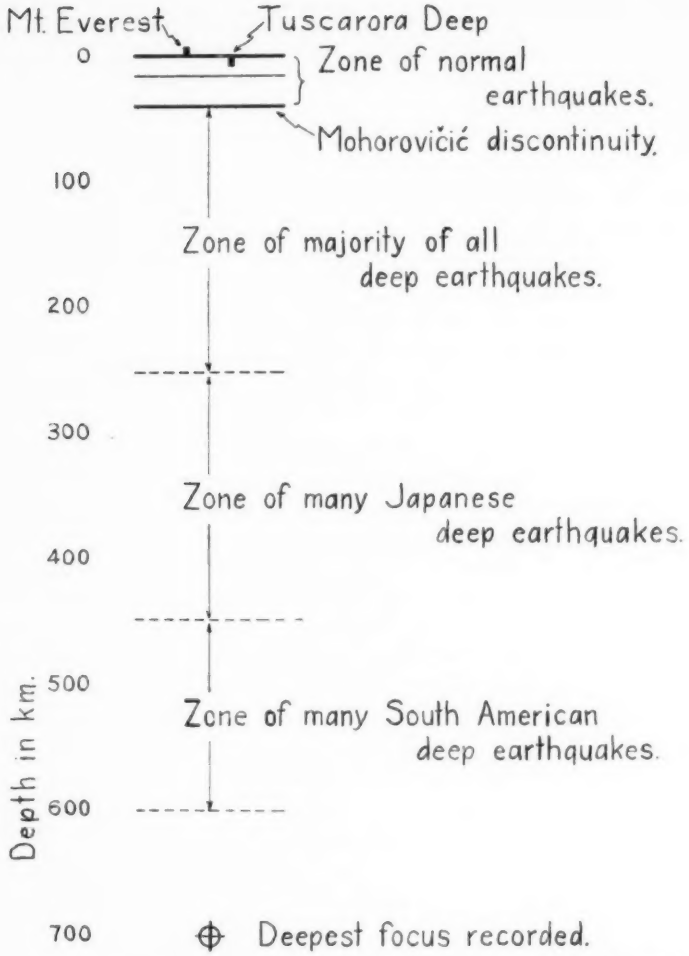

FIG. 1.—Comparison between the elevations of irregularities on the surface of the earth, the thickness of the crust, and the depths at which deep earthquakes are observed to occur.

the velocity of seismic waves and by the results of laboratory experiments upon the compressibility of substances. Within these limitations many density-depth curves have been proposed, an average[9] of which shows the density increasing from about 3.0 in the crust to about 6.5 just outside the core, jumping to 10.5 inside the core, and gradually increasing to about 12.5 at the center of the earth. Based on this distribution of density, the variation of pressure with depth can be derived. At 400 km., the pressure is of the order of $1.5 \cdot 10^{11}$ dynes/cm.^{2}, or 150,000 atmospheres.[10]

RIGIDITY AND BULK MODULUS

The rigidity and bulk modulus can be computed from the density and the seismically determined wave velocities by means of the velocity formulas which have been given. At 400 km. the rigidity is about 10^{12} dynes/cm.^{2}, and the bulk modulus about $1.7 \cdot 10^{12}$ dynes/cm^2.

TEMPERATURE

The distribution of temperature within the earth cannot be determined without making a great many assumptions, so that the thermal condition is consequently one of the most uncertain of the properties of the earth's interior. In the computation of the temperature, the following quantities are required: the temperature gradient near the surface, which has been measured at relatively few places, and these only on the continents; the thermal conductivity, which must be extrapolated from surface values; the amount and depth distribution of radioactive materials, which can only be guessed at; and the age and original thermal condition of the earth. Owing to the latitude available in the choice of numerical values for the various quantities, there is naturally a wide variation in the estimates of internal temperature, ranging from Jeffreys' value of 2,500° C. for the temperature at a depth of 400 km. to DeLury's estimate of 1,500° C., or less, at the same depth.[11]

[9] B. Gutenberg, *Aufbau der Erde* (Berlin: Gebrüder Borntraeger, 1925).

[10] *Ibid.*, p. 51.

[11] H. Jeffreys, *The Earth* (2d ed.), p. 154; J. S. DeLury, *Jour. Geol.*, Vol. XLI (1933), p. 755.

VISCOSITY

The "viscosity" of a material is its internal resistance to plastic flow, or defined more quantitatively, it is the ratio of stress-difference to twice the rate of shear during plastic flow. Experimentally it is observed that the viscosity of a material is increased by increased containing pressure, and decreased by increased temperature.

The distinction between viscosity and rigidity should be emphasized. Rigidity is the ratio of shearing stress to shearing strain for stresses of such a magnitude and rate of application that there is an absence of appreciable flowage. Generally speaking, the higher the viscosity of a material, the higher the rigidity.

A few quantitative estimates of the viscosity of the mantle of the earth have been made. Jeffreys,[12] basing his computation on the existence of the variation of latitude, estimates the viscosity of the mantle to be greater than $5 \cdot 10^{21}$ dyne-cm.-sec. Haskell,[13] applying his solution of the problem of the motion of a viscous fluid under a surface load to the data for the rise of Fennoscandia after the melting of the ice sheet, obtained a value of 10^{22} dyne-cm.-sec. It is interesting to note that this value is about a thousand times as great as the viscosity of a very solid material with which we are familiar—steel.

STRENGTH

The strength of a material is defined as that critical stress-difference above which the rate of change of shape does not decrease when the time of application of the stress increases, or below which the rate of change of shape approaches zero as the time of stress application increases indefinitely. The strength of most materials is increased by pressure and decreased by higher temperatures.

On the important subject of the strength of the material within the earth there is the widest possible divergence of opinion. For example, Daly[14] believes that the material a short distance below the crust has zero strength, although high viscosity; i.e., that the material will respond to differential stress by flowing, however slight

[12] *Op. cit.*, p. 267.

[13] A. N. Haskell, *Physics*, Vol. VI (1935), p. 265; *ibid.*, Vol. VII (1936), p. 56.

[14] R. A. Daly, *Igneous Rocks and the Depths of the Earth* (New York: Mcgraw-Hill, 1933), p. 234.

the stress-difference. Jeffreys[15] takes a slightly different position. He believes that strength may increase in the crust until it becomes several times its surface value, then decrease until it becomes at 50 km. perhaps one-tenth that at the surface, and that strength may just exist at 300 km. T. C. Chamberlin, on the other hand, expressed the view that at least the outer seven-eighths of the earth's volume possesses elastic solidity and argued strongly against the existence of a liquid substratum.[16]

CHARACTERISTICS OF DEEP-FOCUS EARTHQUAKES

DISTRIBUTION

Although earthquakes have occurred nearly everywhere on the earth, the vast majority are at present confined to two great belts, one extending from the tip of South America around the Pacific to New Zealand and the other extending from Portugal eastward and intersecting the first one in the neighborhood of Formosa. Deep-focus earthquakes of depth exceeding 100 km. are generally restricted (within the relatively short period of observation) to four limited regions within these great belts of seismic activity: in South and Central America, in the neighborhood of Japan, in the East Indies, and in the Himalayas.

The distribution of individual epicenters for three of these regions is shown in Figures 2, 3, and 4, on which are plotted the epicenters of earthquakes which are reliably known to have had depths in excess of 100 km., as well as epicenters of the major shallow earthquakes which have occurred over a period of several years.[17]

[15] *Op. cit.*, pp. 185 and 202.

[16] *Jour. Geol.*, Vol. XXXV (1927), pp. 89–94.

[17] These epicenters are plotted from the authors' list of 135 earthquakes of depth exceeding 100 km. which occurred during the years 1919–29, 1932–34, with a few cases for 1930–31. The list was prepared from the *International Seismological Summary*, the preliminary bulletins of the Jesuit Seismological Association, and from various papers on the subject of deep-focus earthquakes. Care was taken to include in the list only those epicenters for which it was felt that enough data had been utilized to make the determination of depth reasonably certain. Epicenters of earthquakes for which the depth was abnormal, but less than 100 km., were not included, since until recently these have been difficult to identify with certainty. Epicenters of the shallow earthquakes were taken from the *International Seismological Summary*, an earthquake being arbitrarily called "major" if its *P* phase had been recorded an epicentral distance greater than 85°.

Approximately 90 per cent of all earthquakes deeper than 100 km. and 65 per cent of all major normal earthquakes occur in the regions represented by these three maps. No map has been prepared for

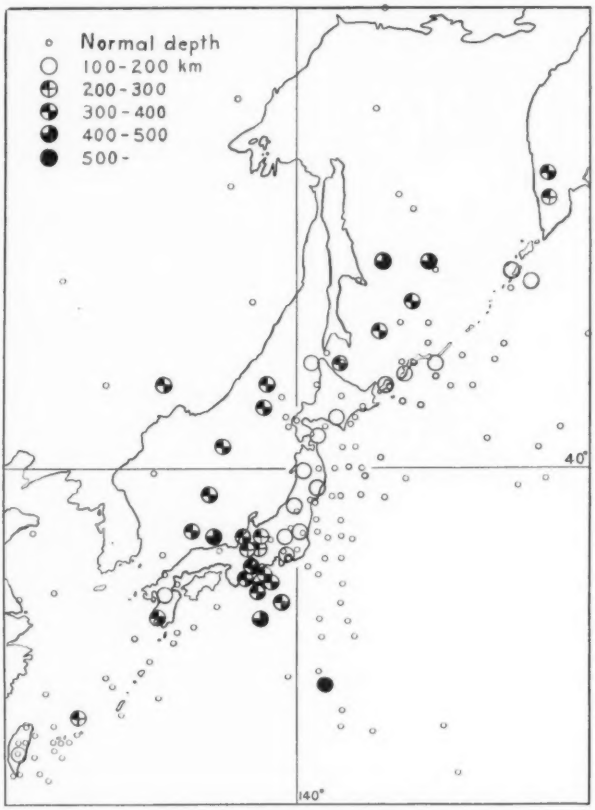

FIG. 2.—Earthquake epicenters in the Japanese region. Deep earthquakes of 1919–34, and major shallow earthquakes of 1924–29.

the Himalayan region, owing to the small number of only moderately deep earthquakes which have occurred there.

In the Japanese region (Fig. 2) there appear to be two systems of

earthquakes of depth greater than 200 km., one extending southwest from Kamchatka through the Sea of Okhotsk and the Japan Sea nearly to Formosa, and the other at right angles to the arc of the islands, extending from about 1,500 km. at sea, across Japan and the Japan Sea, and possibly a short distance into Siberia. East of the first-named line is a belt of intermediate depth, 100–200 km. This zone coincides more or less with the line of the islands. Still farther east is the zone of most frequent occurrence of shallow earthquakes. It should be noted that there are relatively few epicenters of shallow earthquakes in the interior of Japan and still fewer in the Japan Sea. These features of the distribution of Japanese earthquakes were first pointed out by Wadati and Isigawa.[18]

In the South and Central American region (Fig. 3) the earthquakes with depths exceeding 200 km. lie, with two exceptions, inland. There appear to be two lines of earthquakes of intermediate depth, trailing off northwestward and southwestward from the oval zone enclosing the deeper earthquakes. As in the Japanese region, the average depth appears to increase with the distance inland, and the number of shallow earthquakes to decrease.

In the East Indian region (Fig. 4) we note that the deep earthquakes occur in several scattered groups, and there seems to be no system discernible in the distribution of the earthquakes of different depths.

To summarize the distribution noted we can say that only a small part, perhaps a third, of the great belts of seismic activity have associated with them regions of deep-focus earthquakes. The zones of deep-focus earthquakes do not necessarily coincide with the zones of highest frequency of shallow earthquakes, and, in fact, there appears to be, in the Japanese and South American regions, a progression to earthquakes of successively deeper foci in passing inland from the Pacific Ocean.

It should be noted that the period of observation of deep-focus earthquakes is comparatively short—some fifteen years. Hence any conclusions based on the distribution of epicenters as indicated at

[18] K. Wadati and T. Isigawa, *Geophys. Mag.*, Vol. VII (1933), pp. 291–305.

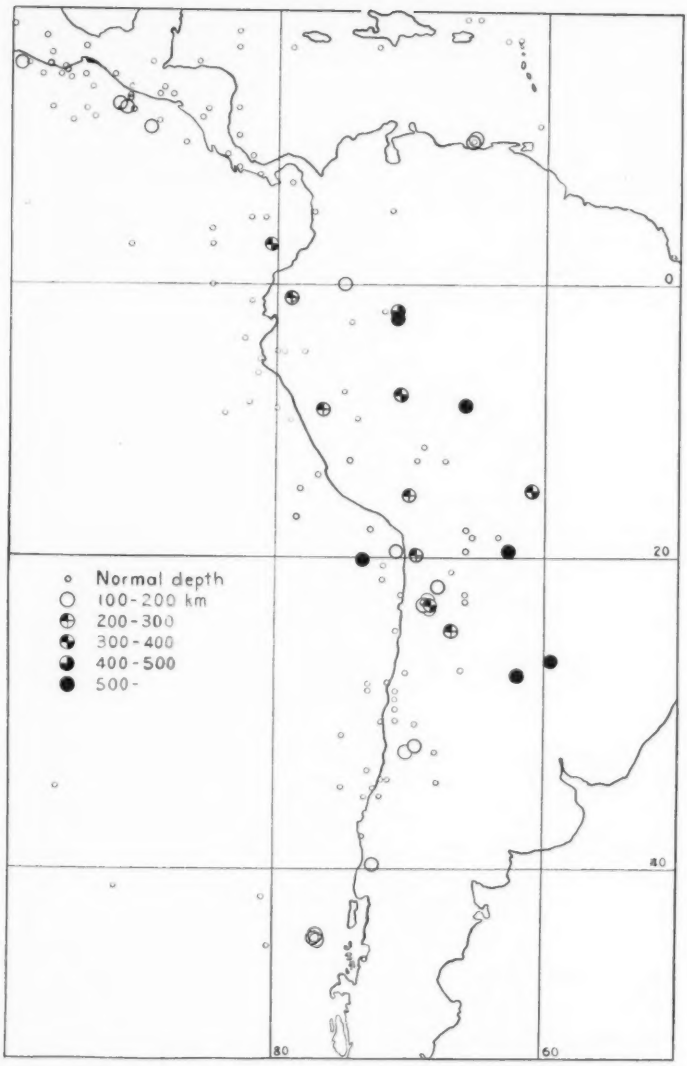

FIG. 3.—Earthquake epicenters in the South and Central American region. Deep earthquakes of 1919–29, 1932–34, and major shallow earthquakes of 1924–29.

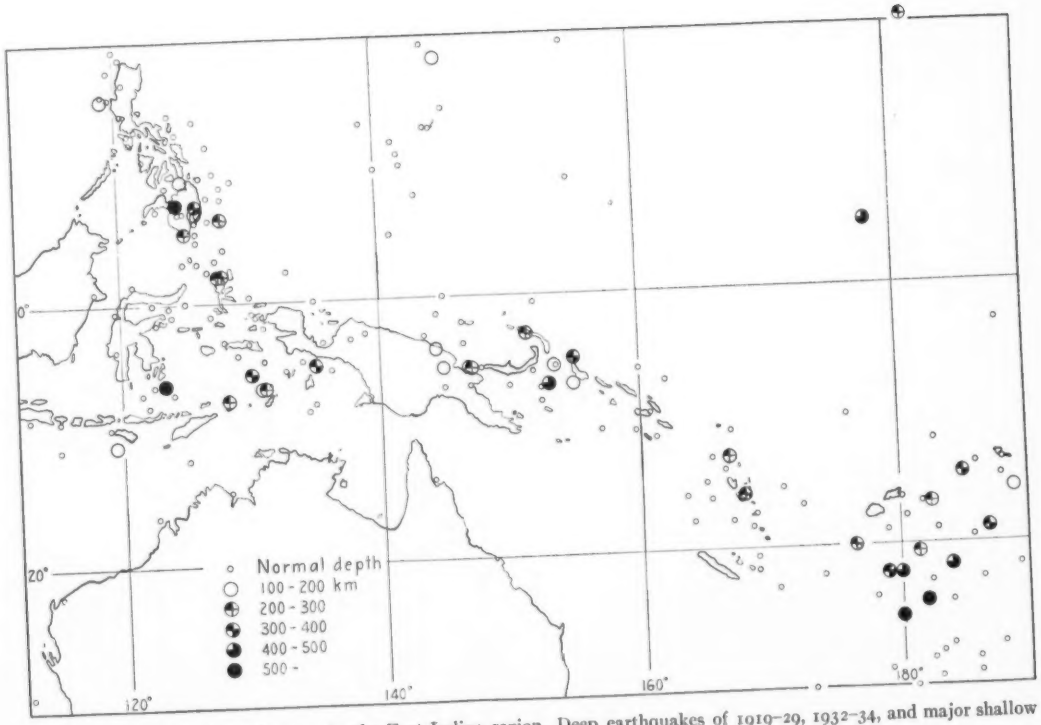

FIG. 4.—Earthquake epicenters in the East Indian region. Deep earthquakes of 1919–29, 1932–34, and major shallow earthquakes of 1927–29.

present are admittedly tentative and may need to be modified if additional years of observation show any significant changes in the distribution.

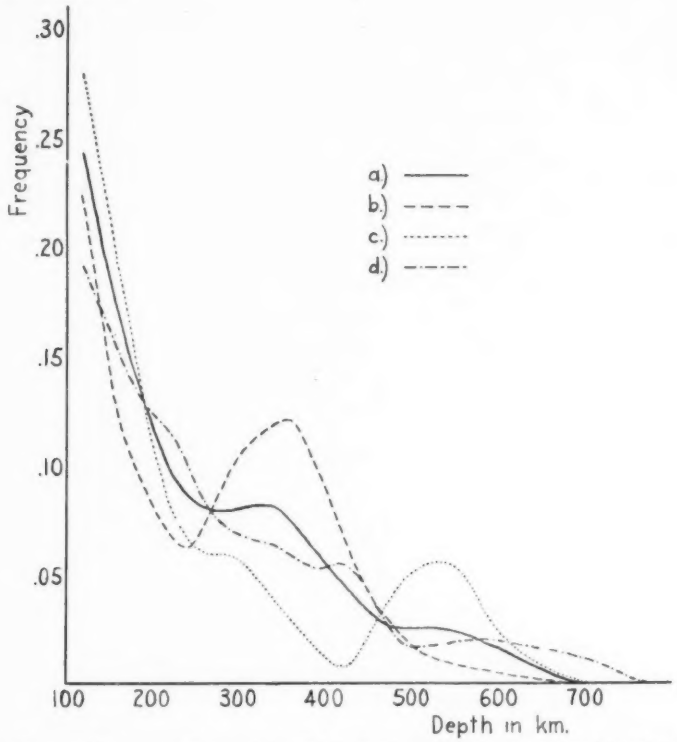

FIG. 5.—Frequency of occurrence of deep-focus earthquakes, as a function of focal depth: (*a*) for all regions, 135 foci; (*b*) for the Japanese region, 45 foci; (*c*) for the South and Central American region, 37 foci; (*d*) for the East Indian region, 46 foci.

DEPTH DISTRIBUTION

Figure 5 indicates how the frequency of occurrence of deep-focus earthquakes varies with depth. The figure includes a depth-frequency curve for all the deep earthquakes on our list and for each of the three major regions. They were prepared from running means of the number of earthquakes in successive 40-km. intervals.

Curve *a*, giving the variation in frequency with depth for the entire list of 135 earthquakes, shows the frequency dropping off in a roughly exponential fashion with increasing depth. Curve *b*, for the Japanese region, shows the frequency rapidly dropping off initially but rising again into a well-developed secondary maximum between the depths of 300–400 km. Curve *c*, for the South and Central American region, likewise shows a definite secondary maximum, but for depths of 475–575 km. Curve *d*, for the East Indian region, shows no definite maxima and is rather similar to the depth-frequency diagram for the complete list of deep earthquakes.

Again it seems advisable to call attention to the short period of observation and the possibility that a fair indication of long-time conditions has not been shown. The maxima here noted may be of some significance, but it should be emphasized that an increase in the number of observations may prove that the depth distribution presented here is only a statistical fluctuation.

ENERGY RELEASED

Shallow earthquakes capable of doing damage release energies in excess of about 10^{15} ergs. For example, the destructive Long Beach, California, shock of March 10, 1933, had an energy of the order of 10^{18} ergs.[19] Deep-focus earthquakes may release as much energy as the great earthquakes of normal depth,[20] although, of course, they are not destructive, owing to the large distance from the focal region to the surface of the earth. Kawasumi calculated that the earthquakes of June 2, 1931, depth 250 km., and June 2, 1929, depth 300 km., released energies of the order of 10^{21} ergs,[21] although these were certainly not the most intense deep-focus shocks that have occurred.

In so far as the question of the range of energy which may be released in deep-focus earthquakes is concerned, the opinion of seismologists who have examined large numbers of seismograms of these earthquakes is that the range is comparable with that shown by normal earthquakes, i.e., from negligible amounts of energy to

[19] C. F. Richter, *Seis. Soc. Amer. Bull. 25* (1935), pp. 1–32.

[20] Wadati and Isigawa, *loc. cit.*

[21] H. Kawasumi, *Ethq. Res. Inst. Bull. 11* (1933), p. 450; *ibid., Bull. 12* (1934), p. 682.

energies of the order of 10^{24} ergs. This impression is verified by an examination of the variation in the number of stations reporting the *P* phase of deep-focus shocks in the *International Seismological Summary*, ranging from half a dozen to over one hundred.

AFTERSHOCKS

For months following a major normal earthquake less intense shocks occur in the same epicentral region, with both the intensity and the frequency of these aftershocks decreasing with the lapse of time after the main earthquake. The total number of aftershocks to a large shallow earthquake frequently exceeds one thousand. In contrast to this is the situation concerning deep-focus earthquakes, following which aftershocks are rarely observed.[22] A search by the authors through the *International Seismological Summary* for the years 1919–29 has yielded only twenty cases of possible aftershocks from foci deeper than 75 km. Nineteen of these followed earthquakes of depth 75–100 km., and only one, whose identification was doubtful, followed an earthquake whose focus was at 175 km. During the period August, 1933—January, 1935, there were five strong deep-focus earthquakes in the South American region at depths ranging from 160 to 350 km. There is inconclusive evidence for the occurrence of an aftershock following one of these (December 4, 1934) which occurred at the depth of 200 km.[23] The evidence so far seems to justify a tentative conclusion that the relative intensity and frequency of aftershocks decrease rapidly with increasing depth of focus, becoming very small at depths not much greater than 200 km.

PERIODICITIES AND CORRELATIONS

For many years attempts have been made to discover periodicities in the frequency of occurrence of shallow earthquakes and correlations between the occurrence of earthquakes and other natural phenomena. No great success has followed such attempts. Recently, similar attempts have been made for deep earthquakes with, in general, somewhat more success.

Conrad[24] has analyzed for periodicities all the deep foci listed in

[22] Wadati, *op. cit.*, Vol. IV (1931), p. 245.

[23] N. H. Heck, personal communication.

[24] V. Conrad, *Gerl. Beitr. 40* (1933), pp. 113–33.

the *International Seismological Summary* for the interval 1918–28. He finds evidence for six- and fourteen-month periods, which he associates with the harmonic components of the rotation of the earth's axis (variation of latitude). The authors have likewise found good evidence for the existence of the fourteen-month period in their more critically selected list of deep foci. Stetson[25] has explored our list for evidence of lunar periodicities in the frequency of deep-focus earthquakes and finds an excellent correlation between the frequency of occurrence and the magnitude of the lunar tidal force at the epicenter at the time of origin of the earthquake. Stetson[26] has also analyzed the major shallow earthquakes which have occurred over a considerable period of years. The evidence for lunar periodicities was much less definite than in the case of deep-focus earthquakes. However, the shallow earthquakes from a limited, structurally homogeneous region have been recently shown to have a good correlation with the lunar semidiurnal period.[27] Yamaguti[28] presents statistical evidence which indicates that deep earthquakes are more likely to occur where the barometric pressure gradient is great than where the barometric pressure itself is great, and also that the depth of focus depends in a linear fashion on the magnitude of the pressure gradient.

Thus there appear to be correlations between the occurrence of deep-focus earthquakes and barometric conditions, lunar tidal forces, and the variation of latitude. It is not to be thought that the exceedingly minute stresses associated with any of these phenomena are the primary causes of deep earthquakes, but rather that these stresses affect the frequency through a triggering action. In addition, any conclusions based on the correlations between the occurrence of deep-focus earthquakes and other phenomena should be recognizedly tentative, since the analysis of a larger amount of statistical material, when it becomes available, may prove these correlations to have been entirely fortuitous.

[25] H. T. Stetson, *Science*, Vol. LXXXII (1935), pp. 523–24.

[26] Personal communication.

[27] M. W. Allen, *Seis. Soc. Amer. Bull. 26* (1936), p. 147.

[28] S. Yamaguti, *Ethq. Res. Inst. Bull. 12* (1934), p. 752.

PATTERNS OF INITIAL MOTION AT THE SURFACE

The principal phases in the seismogram of a deep-focus earthquake are characteristically sharp and clear, making it practicable to determine the direction of initial motion of the surface of the earth. For the compressional (*P*) phase, the initial motion is directed either toward or away from the epicenter, depending on whether the initial pulse in the direction of the station under consideration is one of rarefaction or of compression. When the direction of first motion at a sufficiently large number of stations recording a given earthquake is plotted on a map, a systematic pattern of direction of initial motion is recognizable. The existence of these patterns results from the fact that in certain areas on the map all the first motions are toward the focus, whereas in adjacent areas all the first motions are away from the focus. The lines separating these areas are called "nodal curves." At any point on a nodal curve the magnitude of the motion is zero.

Five of the most common patterns of the direction of initial motion in the compressional phase are illustrated in Figures 6–10.[29] The scale chosen is approximately that for an earthquake at a depth of 250 km., although the size of the focal region is greatly exaggerated. In each of these figures, *a* represents a typical pattern of first motion as observed at the surface and *b* represents a qualitative description of a process at the focus equivalent to the one which produced the given surface pattern. In each case it appears that the process, whatever it may actually be, can be likened to the transformation of a sphere into an ellipsoid which contains, in the case of an ellipsoid of revolution, two cones whose generating lines are lines of no distortion, i.e., lines in the direction of which no change in length has occurred. If the transformation has been from a sphere to a triaxial ellipsoid, there will be, instead of cones of no distortion, two

[29] Numerous maps of observations of directions of initial surface motion are to be seen in the following papers: H. Honda, *Geophys. Mag.*, Vol. IV (1931), p. 185; *ibid.*, Vol. V, pp. 69–88; *ibid.*, Vol. VI (1932), pp. 301–26; *ibid.*, Vol. VII (1933), pp. 257–68; Y. Oka, *ibid.*, Vol. VI (1932), pp. 213–22; K. Hayata, *ibid.*, pp. 363–69; M. Ishimoto, *Ethq. Res. Inst. Bull. 10* (1932), pp. 449–71; *ibid.*, *Bull. 11* (1933), pp. 254–73; T. Fukutomi, *ibid.*, pp. 510–29; H. Kawasumi, *ibid.*, Vol. XI, pp. 403–53; *ibid.*, *Bull. 12* (1934), pp. 660–705; H. Kawasumi and R. Yosiyama, *Proc. Imp. Acad. Jap.*, Vol. X (1934), pp. 345–48.

planes of no distortion, each being a circular section of the original sphere.

Consider first the surface pattern *a* of Figure 6, for which the motion is away from the epicenter within a small nodal circle sur-

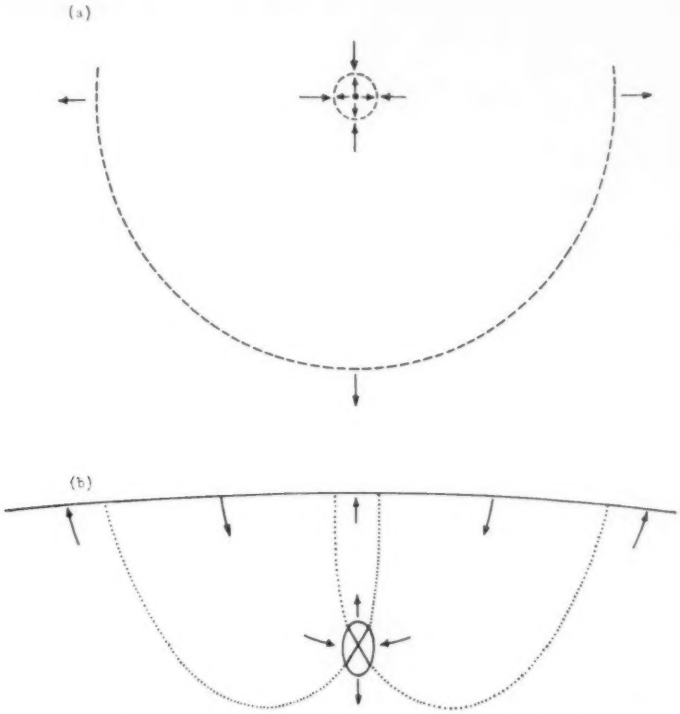

FIG. 6.—(*a*) Surface pattern of the direction of first motion in the *P* phase; (*b*) its equivalent focal ellipsoid (prolate).

rounding the epicenter, toward the epicenter between this circle and a larger nodal circle, and away from the epicenter everywhere outside the larger circle. The process in the focal region which produces this surface pattern is equivalent to the passage of a sphere into a prolate ellipsoid whose long axis is vertical, and which possesses two

cones whose generating lines are lines of no distortion (*b*, Fig. 6). The scale to which this ellipsoid is drawn is of no significance. Within the small nodal circle surrounding the epicenter, the motion is away from the focus, corresponding to the concept that within the cone the radii of the original sphere have been lengthened. Outside the small nodal circle and within the large one the motion is toward

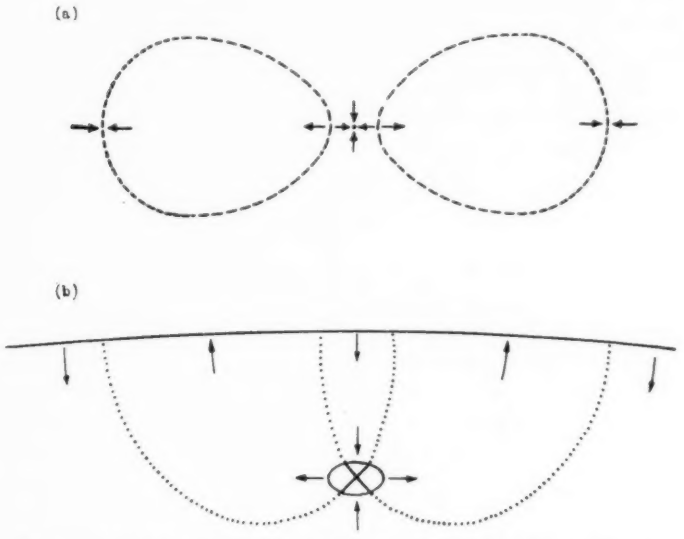

FIG. 7.—(*a*) Surface pattern of the direction of first motion in the *P* phase; (*b*) its equivalent focal ellipsoid (prolate).

the focus, corresponding to the shortening of the radii of the sphere outside the cones. The motion outside the large nodal circle is similarly accounted for by the lower cone.

The patterns of first motion shown in Figures 7*a* and 8*a* can also be accounted for by the passage of a sphere into a prolate ellipsoid. In Figure 7 the long axis of the ellipsoid is horizontal, and in Figure 8 the long axis is inclined 45° (assuming that the included angle of the cones is 90°).

In order to account for the patterns shown in Figures 9 and 10,

the equivalent process which must be postulated is the passage of a sphere into a triaxial ellipsoid, whose intermediate axis has the same length as the diameter of the sphere. Such an ellipsoid has two circular sections, with the same radii as the sphere, as planes of no distortion. The pattern shown in Figure 9 can be qualitatively ac-

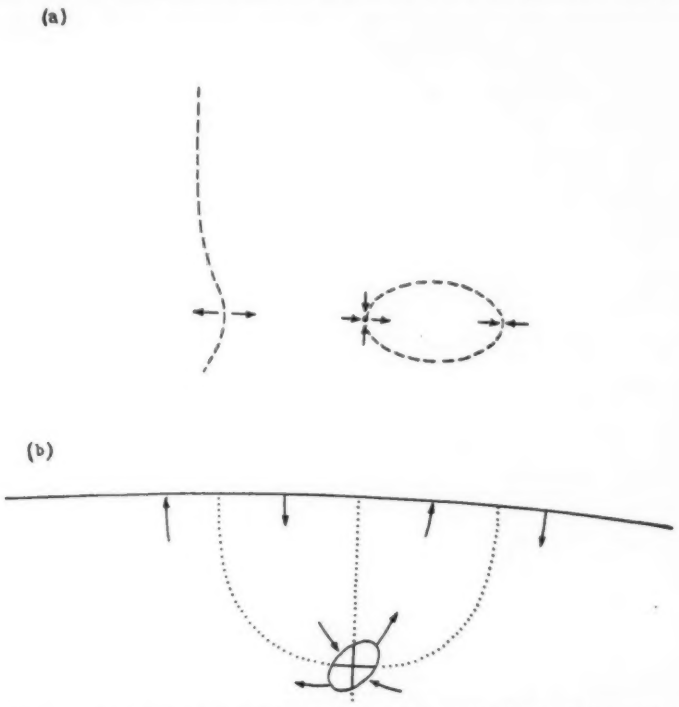

FIG. 8.—(*a*) Surface pattern of the direction of first motion in the *P* phase; (*b*) its equivalent focal ellipsoid (prolate).

counted for by a process at the focus equivalent to the transformation of a sphere into a triaxial ellipsoid whose intermediate axis is horizontal and whose long axis is inclined 45° (again assuming the angle between the surfaces of no distortion to be 90°). Similarly, Figure 10 can be explained by a triaxial ellipsoid whose intermediate axis is horizontal and whose long axis is vertical.

Not only the pattern of direction of initial motion in the *P* phase but also the pattern of first motion in the *S* phase can be accounted for qualitatively by assuming that the process at the focus is equivalent to the transformation of a sphere into an ellipsoid. Generally speaking, the nodal lines for the *P* phase are lines of maximum ampli-

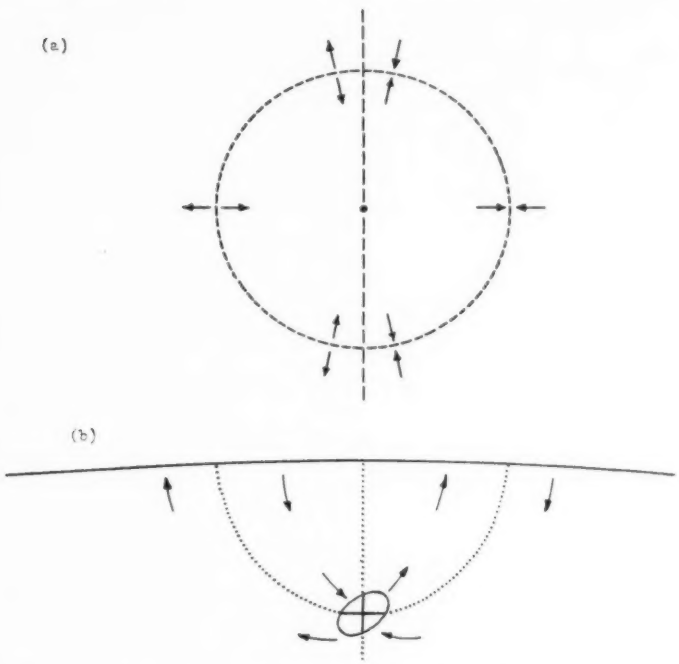

FIG. 9.—(*a*) Surface pattern of the direction of first motion in the *P* phase; (*b*) its equivalent focal ellipsoid (tri-axial).

tude of the *S* phase, corresponding to the fact that cones or planes of no distortion may likewise be surfaces of maximum shear. The patterns of direction of first motion in the *S* phase have nodal curves of their own, these coinciding generally with the lines of maximum amplitude of the *P* phase, which in turn signify directions in the ellipsoids of maximum elongation and zero shear.

All the patterns of initial motion in the *P* phase which have been observed correspond to one of the types illustrated in Figures 6–10, with slight variations due to slightly differing inclinations of the axes of the ellipsoids. Of the various types, those illustrated by Figures 6, 7, and 9 are by far the more common. A pattern which might be described by the transformation of a sphere into an oblate

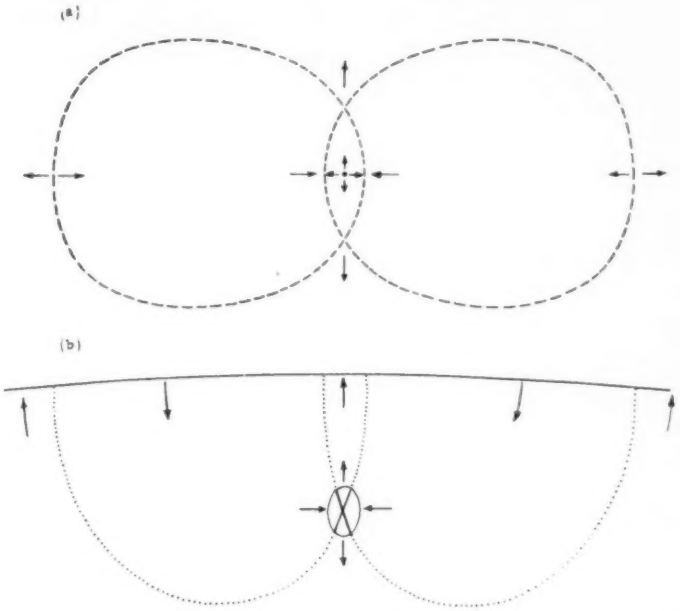

FIG. 10.—(*a*) Surface pattern of the direction of first motion in the *P* phase; (*b*) its equivalent focal ellipsoid (tri-axial).

ellipsoid has never been observed. Of particular importance, when the significance of deep-focus earthquakes comes to be discussed, is the fact that no earthquake has been observed for which the initial surface motion was everywhere away from, or everywhere toward, the focus. That is, in the processes resulting in earthquakes, there are invariably compressions produced in some directions, and rarefactions in others.

It should be emphasized that the actual physical process at the focal region resulting in the liberation of seismic waves is not necessarily the passage of a spherical region into an ellipsoidal region, but that the observations of patterns of initial motion which have been described here merely indicate that, whatever the actual process at the focus may be, it is, in producing elastic waves, equivalent to such a transformation.

The foregoing discussion of the patterns of initial motion applies to shallow as well as to deep earthquakes. However, clearly defined patterns are much more difficult to observe in the case of shallow earthquakes, owing to the generally more complicated nature of the waves, the effects of crustal discontinuities on the waves, and the necessity of having a much denser network of observing stations.

STRESS AND STRAIN NEAR THE FOCAL REGION[30]

When the distribution of amplitude of initial motion, in addition to direction, is utilized, it becomes possible to obtain the displacement, stress, and strain set up by the passage of the elastic waves produced by an earthquake.[31] From the location of the focus, the pattern of direction of initial motion, and the known variation of the velocity of elastic waves with depth, one can determine the type of equivalent focal ellipsoid, the orientation of its axes, and the angle included between its surfaces of no distortion. Next the appropriate solid spherical harmonic solution of the equation of propagation of elastic waves which changes signs at the polar angles corresponding to the cones or planes of no distortion is selected. This function will give the observed pattern of both the direction and the magnitude of initial surface motion—the latter on an arbitrary scale owing to the presence of an arbitrary constant multiplier in the solution. After the determination of the value of this constant by comparison of the values of surface motion predicted by the function with the observed values, the function is available for the determina-

[30] The material of this section, including the derivation of the expression for the maximum radius of the focal region, was given by one of the authors (J. A. S.) at the April, 1936, meeting of the Seismological Society of America.

[31] Honda, *op. cit.*, Vol. V (1932), pp. 301–26; *ibid.*, Vol. VII (1933), pp. 257–68; Kawasumi, *op. cit.* (1933), pp. 403–53; *ibid.*, *Bull. 12* (1934), pp. 660–705.

tion of the displacement, stress, and strain in the elastic waves at any desired point in the interior of the earth.

Since the function just described is derived as a solution of the equation of motion of an elastic medium, it is inapplicable within the focal region, where the elastic limit of the material has obviously been exceeded. Consequently, if we can get an estimate of the maximum radius of the focal region, we can set a lower limit to the value of stress which must have existed within the focal region during or immediately before the earthquake. It can be shown that the radius of the focal region is always less than

$$r = 2.5V\left(\frac{\Delta T_s}{\Delta T_p} - 1\right)\Delta T_p ,$$

in which V is the average velocity of compressional waves to the depth of the focus, ΔT_p is the duration of the initial compressional pulse recorded at the surface, and ΔT_s is the duration of the initial transverse pulse.

The numerical results obtained by Kawasumi[32] in his study of the earthquake of June 2, 1931, will be briefly summarized. From the travel-times of the P and S waves, the epicenter was found, and the depth of focus determined as 250 km. The pattern of direction of first motion in the P phase indicated that the equivalent focal ellipsoid was of the prolate type, with its principal axis directed southwest and dipping 67°, and that the included angle of the cones was 110°. The appropriate mathematical function was chosen and its arbitrary constant determined, and the function then used to derive values of displacement, compressional and shearing stress, and strain, as a function of polar angle and of distance from the center of the focal region. Kawasumi records that the periods of the waves as read from the seismograms at one station were 2.7 sec. for both the transverse and the compressional waves, which implies, according to our formula, zero radius of the focal region. Thus we shall be on the safe side by assuming that the ratio of the duration of the transverse pulse to that of the compressional pulse was somewhat larger than this, say 1.15:1. Using this as the value of $\Delta T_s/\Delta T_p$

[32] Kawasumi, *op. cit.*, pp. 403–53.

in the formula, with $\Delta T_p = 2$ sec., and with the value of the velocity appropriate to the depth of 250 km., we get 6 km. as the radius of the focal region. At a distance of 6 km. from the center of the focal region, Kawasumi's curves indicate that the displacement in the elastic waves was about 1 cm., the maximum shearing and compressive stresses of the order of $2 \cdot 10^7$ dynes/cm.^{2}, and the strains of the order of $2 \cdot 15^5$.

SUMMARY: ENVIRONMENT AND CHARACTERISTICS OF DEEP-FOCUS EARTHQUAKES

1. About 10 per cent of all major earthquakes occur below the crust of the earth at depths ranging up to 700 km.

2. The zones of occurrence of deep-focus earthquakes are associated with portions of the great belts of normal earthquakes. For some of the zones there appears to be a progressive deepening of foci with increasing distance inland from the oceanic margins.

3. The number of earthquakes occurring decreases rapidly with increasing depth of focus, although there are, possibly, secondary maxima in the frequency at different depths in different regions.

4. Aftershocks decrease in relative intensity and number with increasing focal depth.

5. There appears to be as wide an energy range for deep earthquakes as for shallow ones.

6. Correlations exist between the periodicity of occurrence of deep-focus earthquakes and the fourteen-month period of variation of latitude and between the occurrence of deep earthquakes and the magnitude of lunar tidal forces. There is evidence for correlation between the epicenters of deep-focus earthquakes and the location of regions of maximum barometric gradients and between the depth of focus and the magnitude of the barometric gradient. All such correlations are, in general, more definite for deep than for shallow earthquakes.

7. Patterns of initial surface motion indicate that in the processes resulting in earthquakes there are invariably compressions produced in some directions and rarefactions in others. The magnitude and directions of the elastic waves produced by the earthquake can be accounted for, in every case so far observed, by assuming that the

mechanism of the process is equivalent to the transformation of a sphere into a prolate ellipsoid or into a triaxial ellipsoid whose intermediate axis remains unchanged in length. The maximum radius of the focal region can be determined, and a lower limit placed on the stress, strain, and displacement within the focal region. For one earthquake investigated the shearing strain exceeded $2 \cdot 10^7$ dynes/cm^2.

8. The absence of first-order velocity discontinuities within the range of depth of deep-focus earthquakes suggests that there are no sudden changes in the density, rigidity, and bulk modulus of the material over this range. These factors, and also pressure and temperature, increase gradually with depth. It is likely that the composition of the material likewise changes only gradually over this depth.

9. Thermal conditions within the earth, either the temperature gradient or the temperature, are not known with any certainty.

10. The viscosity of the material of the mantle of the earth, according to computations based on the assumption that the material is a viscous fluid, is about one thousand times that of steel.

11. The magnitude of the strength of the material of the mantle is unknown.

LOGICAL IMPLICATIONS OF THE DATA PRESENTED

Some of the facts of observation so far presented are unequivocal in their implications as to the nature of the material beneath the crust and its manner of behavior under certain conditions. Other data, however, are subject to more than one interpretation, particularly when analyzed individually rather than collectively. For this reason we shall now attempt to evaluate these data to find their collective bearing upon a few problems of major importance; first, the type of process which results in deep-focus earthquakes; second, the possible sources of stresses which might cause the process to occur; and, third, the strength and probable manner of failure of the substratum under postulated conditions.

TYPE OF PROCESS

The first postulate to be tested is that deep-focus earthquakes are the result of the same sort of fracturing which is responsible for shal-

low earthquakes. In support of this are all the facts and implications which show a consanguinity between the two types. A common factor is indicated by their geographical distribution, by their association with the same natural phenomena in frequency periodicities, and by the fact that both shallow and deep earthquakes produce the same patterns of surface motion and are relatable to the same sorts of non-spherical strain patterns at the foci. The departures from exact geographical coincidence are such as might be expected, with the zones of failure dipping downward under the continents.

In other characteristics, in which the two types of earthquakes differ, the disagreements are such as are easily explained, and are, perhaps, predictable from a sound analysis of the available data. Take, for example, the great difference in frequency. Certainly the greatly increased pressures and the higher elastic constants below the crust should make rupture more difficult than within the crust. In addition, the absence of major discontinuities within the zone of deep earthquakes tells us that changes in the composition and physical properties are gradual. Contrasted with this deep zone, which is denser, more rigid, under tremendous containing pressures, and comparatively homogeneous, is the crust. Actual observation shows us that at least a part of the crust is not homogeneous. It contains innumerable zones of weakness and surfaces of potential failure. The containing pressures are less. Aftershocks follow earthquakes in the crust, but they follow deep earthquakes infrequently. If the same process—fracturing—is active in both zones, it seems evident that the environment of deep earthquakes is such as to heal fractures quickly and completely so that the multitude of residual zones of weakness such as we have in the crust are absent not far below the crust. Thus, the very differences between shallow and deep earthquakes serve to emphasize their consanguinity by their complete agreement with what we know of the two environments.

Let us make an alternative postulate: that the process which results in deep-focus earthquakes is a change in state due to some change in environment. Most of our observations of geographical distribution, depth frequency, energy release, and aftershocks are not incompatible with this postulate. With regard to the periodicities of frequency it is more difficult to find a connection. If the func-

tion of these natural phenomena is merely to change the deep environment by a slight increase in pressure, they seem to be pitifully small. While these periodic stresses might act as triggers to set off a stress system already perilously close to the point of failure, it does not seem possible that their function is only to change an environment by increasing slightly the hydrostatic pressure.

There is another method of analysis which may help us to answer this question of the type of process. Let us suppose that the change of state is of the sort called a "polymorphic transition,"[33] meaning approximately that a given material will change its phase when subjected to a different environment. To result in deep-focus earthquakes with the movement patterns noted, this change in phase must also be a change in the external form of the changing mass of material. The change in external form need not be rapid, but we know it must be rapid enough so that it can build up stresses faster than they can be dissipated in the surrounding medium by flow, or no catastrophic failure would result. We shall consider, then, that the material within the focal region changes form at a relatively rapid rate. The surrounding medium will yield elastically to the changes in dimensions of the zone undergoing the change in state, until its elastic limit is reached, at which point failure will occur in the normal manner, by fracturing. If we consider the change in environment which causes a polymorphic transition to be one requiring external shearing stress, the net effect is that the resulting earthquake would be indistinguishable in any respect from those due to an external stress system not accompanied by a change in state. If, on the other hand, we consider that the change in environment is simply one of changing temperature and changing hydrostatic pressure, there is serious doubt that the ensuing change in state would result in an orientation of the material so as to give a non-spherical strain. The probability is that each unit of the changing mass might acquire a new form, elongating in one direction and shortening in another, but the mass orientation would be adapted to equal pressures from all directions. Thus we tentatively conclude that poly-

[33] P. Bridgman, *The Physics of High Pressures* (New York: Macmillan Co.), chap. viii.

morphic transitions due to a change in environment without the aid of an external stress system do not cause deep earthquakes.

It may also be concluded that deep earthquakes are the result of a fracturing response to external stresses, whether or not accompanied by some polymorphic change. Because of the consanguineous characteristics of deep and shallow earthquakes, the authors favor the postulate that deep earthquakes and shallow earthquakes are the result of ordinary fracturing in response to external stresses, without polymorphic transitions playing an essential part.

POSSIBLE SOURCES OF STRESS

What are the possible sources of external stress-differences which might cause fracturing, resulting in deep earthquakes? The periodic stresses already discussed may play some part if the correlations are valid, but the stresses due to the phenomena of variation in latitude and lunar tides are of the order of 10^4 dynes/cm.^{2}, or less than one one-thousandth of the stresses developed outside of the focal region by an earthquake (10^7 dynes/cm.^{2}). The stresses due to barometric gradients are probably somewhat less than 10^4 dynes/cm.^{2} at the depth of moderately deep earthquakes. We must conclude, therefore, that the periodic stresses which appear to be correlated with deep earthquake frequencies are much too small to be primary causes.

The stresses due to surface irregularities, to the lack of isostatic compensation, must also be dismissed as too small. These can have relatively high values in the crust, of the order of 10^9 dynes/cm.^{2},[34] but they drop off exponentially with depth[35] and become negligible at depths greater than 100 km. Surface irregularities cannot provide sufficient shearing stress at any great depth beneath the crust.

Also owing to their small magnitude, we can dismiss the residual stresses due to recent glacial unloading, which Daly[36] has mentioned as a possible cause of subcrustal earthquakes. In addition to the inadequate order of stress available at depth, there is this objection against this source of stress: that no deep-focus earthquakes have

[34] Jeffreys, *op. cit.* (2d ed.), p. 186.

[35] *Monthly Notices, Royal Astron. Soc., Geophys. Suppl.*, Vol. III (1932), p. 36.

[36] Daly, *op. cit.*, pp. 197–98.

been observed in two of the regions of recent glacial unloading—Fennoscandia and North America.

The equatorial component of the stresses tending to produce continental migration[37] is of the order of 10^4 dynes/cm.^{2}, and is entirely insufficient to cause fracture. This source of stress can also be dismissed on other grounds: that a great many deep earthquakes occur near the Equator, where stresses from this source vanish.

There are a few hypothetical stresses which, if they exist, may be of sufficient size to be considered as primary causes of deep earthquakes. Jeffreys[38] has shown that the contraction produced by the cooling of an originally liquid earth would have produced tangential forces in the crust of the order of 10^9 dynes/cm.^{2} Jeffreys also concludes[39] that if the cooling below the oceans has been 1.5 times the cooling beneath the continents, to a depth of about 400 km., a difference of radial contraction of the order of 10 km. would ensue, implying the production of a shearing stress of the order of 10^8 dynes/cm.^{2} This would be at its maximum at the boundary between ocean and continent and would gradually decrease with depth.

The stresses which would be developed by a slowing-down of the period of rotation of the earth from twelve hours to the present twenty-four hours have been computed by Stonely[40] to be of the order of 10^8 dynes/cm^2.

There is another possible source: thermal convection within the earth. Pekeris[41] has recently solved the problem of the convection within the earth, between the core and the crust, owing to a difference in temperature between the bottom of the oceans and the same level in the continents. His calculations (assuming the viscosity to be 10^{22} c.g.s.) indicate that stresses of the order of 10^7 dynes/cm.^{2} can be produced in the convection, i.e., that convection currents would produce a large viscous drag against the bottom of any shell of ultimate strength greater than 10^7 dynes/cm.^{2}. Thermal convec-

[37] Jeffreys, *op. cit.* (2d ed.), pp. 300 ff. [38] *Ibid.*, p. 285.

[39] H. Jeffreys, *Monthly Notices, Royal Astron. Soc., Geophys. Suppl.*, Vol. I (1924), p. 153.

[40] R. Stonely, *Monthly Notices, Royal Astron. Soc., Geophys. Suppl.*, Vol. I (1924), p. 153.

[41] C. L. Pekeris, *Monthly Notices, Royal Astron. Soc.*, Vol. III, No. 8 (1935), pp. 343–67.

tion, therefore, appears to offer a stress of approximately the order developed in the vicinity of foci by seismic waves.

A consideration of the possible stresses mentioned brings out the following facts: that there are two, or possibly three, very small stresses, of the order of 10^4 dynes/cm.^{2}, all acting, apparently, as effective triggers to set off an already unbalanced stress system. There are other stresses due to the irregular distribution of mass in the crust and on top of the crust which are also very small but which definitely cannot act as triggers. In addition to these stresses there are hypothetical stresses, some of which may be large enough to be considered as primary causes of the stress unbalance. The three large stresses, however, are dependent upon highly speculative factors. A stress dependent upon differential cooling and contraction is also dependent upon the hypothesis of an originally very much hotter earth. It is, also, not easy to reconcile the hypothesis with the periodic nature of earth diastrophism and with the apparent fact that such movements show no signs of diminishing activity through geological history. A stress due to a slowing-down of the earth's speed of rotation relies upon some hypothesis of what the original rate was and upon the age of the earth. Thermal convection appears to be a possibility, provided that the strength of the material does not exceed 10^7 dynes/cm.^{2} (*ca.*) and that the perturbation is not destroyed by circulation.

Whether these stresses, or more probably similar stresses of an as-yet-unknown sort, may be the stresses which keep the subcrust in a constant state of unbalance which is added to from time to time by the small, rapidly acting, periodic stresses is an interesting question. It carries with it the implication that the slow stresses are not being entirely absorbed by plastic flow, for even with a partial flow there is sufficient excess stress to result in an unbalanced condition.

It seems more than likely that none of the stresses here mentioned plays an important part in causing deep earthquakes, and even the largest of these stresses may not exceed the strength of the material below the crust.

STRENGTH

The strength of the material of the earth below its crust is probably the most important question confronting students of diastro-

phism, and, as suggested in the introduction to this paper, the paucity of facts has allowed the greatest latitude of speculative thought. It is the belief of the authors that the problem can be answered definitely by a careful analysis of the data we now have, but first it is necessary to come to a complete understanding of the meaning of certain terms.

An early, and still common, view of strength is that solids possess strength, whereas liquids do not—a satisfying-enough explanation if one can define a solid or a liquid. Unfortunately the dividing line between the two states is vague and in some cases meaningless, particularly with regard to strength. For example, most solids possess rigidity, and water, a liquid, does not. If an elastic solid is stressed, its rate of yield approaches zero as the elastic limit is reached. If, before this limit is reached, the stress is removed, the substance will return to its original shape—as it was before the distortion began. If a liquid is similarly stressed, its rate of yielding does not decrease but continues and may even increase as the time of stress application increases. If the stress is removed, the liquid does not return to its original shape.

Now, if all substances which are solid possessed true elasticity and all liquids did not, the problem would be simple, but some substances possess what may be termed an "imperfect elasticity." An example used by Jeffreys[42] is that of a certain type of biscuit, a gingersnap, which, if bent slightly, will move slowly back to its original shape or at least very close to it. In such a case the "afterworking," as the return to shape is called, is closely allied to the property of rigidity or the tendency to return to an original shape after having been deformed. Such a property may be merely a very low rigidity of an imperfectly elastic substance (as are all substances). The interesting fact is that this imperfect elasticity seems to be a property possessed by non-crystalline substances as well as crystalline.

In addition to the varying degrees of elasticity which different substances possess, it is possible to submit a known crystalline substance to a new environment in which all trace of an elastic limit is lost. Steel is such a substance, which, at sufficiently high temperatures, loses its elasticity and, with it, its strength.

The possession of the property of crystallinity has been used as

[42] H. Jeffreys, *The Earth* (2d ed.), p. 181.

a criterion for separating solids from liquids, but what shall we call ordinary window glass? It is a strong substance in the sense that it is highly elastic and rigid, and it is strong in the sense that it will support slight stresses almost indefinitely without failure, and yet its internal structure is such that it is commonly called an undercooled liquid.

It is apparent that solids and liquids cannot be distinguished upon the basis of elasticity or upon the basis of crystallinity, although no one doubts that solids as a class possess greater strength than liquids as a class. It is equally apparent that even the known phases of solids and liquids are functions of the environment. A substance which is solid at one temperature becomes liquid at another. Since no exact line can be drawn between solid and liquid substances or, perhaps, between strong and weak substances, we are forced to use these terms qualitatively. One substance is relatively stronger than another, or a given substance is relatively stronger in one environment than in another.

Let us now consider the factors which control the relative strength and relative solidity of substances. If we subject a crystalline substance to increasing temperatures, we find that, eventually, the substance reaches a point where its constituent units begin to separate and the mass to expand. The interatomic bonds are weakened through distention, and, if the temperature effect continues, the units become so far separated that the space lattice of the original crystal is lost, and the substance is in what we normally recognize as the liquid state. In a liquid just above the melting point the space lattice, or traces of it, may still exist, but the bonds between the units are definitely weakened and the units have separated. If the effect is continued still further, the dissociation may become so pronounced that the substance is in a gaseous state in which the only intermolecular contact is through collision. In nearly every substance (there are a few exceptions) temperature has the effect of loosening the bonds between the constituent units and expanding the mass as a whole through separating the units. In most substances there are critical points where the change is not gradual but rapid, and it is these points which determine the melting point or the boiling point of a given substance at a given pressure.

There is another factor which must be considered as acting direct-

ly against temperature, and that is pressure. The effect of pressure is to compact the units of a substance, to press them closer together and to prevent the dissociation which increased temperature demands. Thus it is that higher pressures raise the boiling point and the melting point.

Thus there are two opposing factors, temperature and pressure, which control the state of a substance and make it either solid or liquid, as we normally use the terms. These factors are constantly at work regardless of whether the substance is crystalline or amorphous. Even an undercooled liquid, such as brittle tar, is more closely compacted by higher pressure and more loosely held together at higher temperatures. Its strength, relatively speaking, must therefore be a function of the relative effects of these two factors. A substance such as steel grows weaker if the temperature is increased relative to the pressure, until, at "cherry red" it loses most of its strength, and at the melting point it loses most of what is left. Even in the liquid state, as the temperature increases and if the pressure remains constant, its viscosity continues to get less and less, which is another way of saying that its strength is diminishing relatively. We conclude, then, that the strength of a material, regardless of its type of internal structure, is dependent upon the combined effects of temperature and pressure, and the strength or weakness of a material is relative.

With this understanding of the meaning of strength, let us reexamine the data concerning the material below the earth's crust. Below the crust, from the Mohorovičić discontinuity down to the core, the velocities of seismic waves increase steadily with, so far as we know, no reverses. From the formulas for the velocities of seismic waves we know that the velocities depend upon ratios between the rigidity and bulk modulus and the density. Knowing, within very close limits, the density for any given depth and the velocity for the same depth, it can be shown that the rigidity and the incompressibility increase steadily downward. Now, rigidity is a property of substances which is subject to the same controlling factors as strength. For example, a tuning fork which will vibrate so as to give a clear tone can be made from a non-crystalline material, such as sealing wax, at low temperatures. Such a fork possesses very high rigidity, but if the temperature is increased, the rigidity is lost and the wax melts. Acting in exactly the opposite way, increasing

pressures raise the melting point and tend to increase the rigidity and the incompressibility as well.

The velocity of seismic waves is influenced indirectly and the strength of materials is influenced directly by the same factors—temperature and pressure. If the strength of a material is observed to decrease, it is probable that the change in environment which causes the decrease is such that the temperature-pressure ratio of effect increases. If the temperature-pressure ratio can be observed to decrease, it is probable that the strength increases.

In the earth it has been demonstrated that the velocities of seismic waves increase downward. Consequently the resulting effects of pressure and temperatures at increasing depths must be such that pressure is increasingly dominant over temperature, and the strength of the earth must increase as this pressure-temperature ratio increases. From the facts that the rigidity of the material and the velocities of seismic waves increase with depth, and from the fact that strength is dependent upon the same controlling factors, we reach the tentative conclusion that the strength of the earth's material must also increase with increasing depth.

GEOLOGICAL CONCLUSIONS

We have shown evidence that the strength of the material of the earth, from the surface downward at least to the core, increases constantly. What does this conclusion mean with respect to the problem of how the subcrustal material fails during diastrophism?

First of all, it settles unequivocally the possibility of liquid flow. Whatever its true manner of failure may be, it will not respond to the slightest of stresses as will a liquid. From a solid substance, such as granite, the earth grows continually stronger downward, and its inter-unit bonds are strengthened accordingly instead of being weakened as they are in a liquid. In what other manner may it fail, as we know it must fail, to accommodate the known movements of the crust?

It may fail by normal fracture as the crust does. This we know to be true from our analysis of deep-focus earthquakes, which shows beyond doubt that the material down to depths of at least 700 km. fractures in much the same manner as the material immediately at the surface. Each of these deep earthquakes is evidence of some

movement at depth, and it would be well to bear in mind that upward of ten thousand earthquakes are recorded each year, a certain percentage of which are deep seated. The net movement due to all these, over periods of thousands of years, may well be more than enough to account for the observed crustal movements.

But there is another way in which the earth's material may fail other than by normal fracture. Solid, strong materials can yield to differential stress without conspicuous fracture through the process of recrystallization, accompanied sometimes by the processes of mylonitization and gliding. It is the process commonly designated as rock flowage which is not to be confused with the flow of liquids. Just what are the factors of environment which result in this type of yielding?

The classic experiments of Adams, in which marble was made to flow by distorting it under high containing pressures, demonstrated that high pressures favor this type of failure, and that stress differences greater than those necessary to cause fracture when the marble is unsupported are necessary to cause the flow. Increased pressures, therefore, raise the resistance of the material to distortion but at the same time favor the ultimate yield by flow rather than by fracture.

The effect of temperature is similar in that increasing temperatures also favor the type of yield characterized by flow. Through weakening the material and its interunit bonds the material will yield more readily by flow at high temperatures than at low temperatures. This effect is demonstrated whenever a blacksmith heats iron in order to bend it.

Our question as to what conditions result in rock flow seems again to depend upon the evaluation of temperature and pressure conditions, but in this case we find that both temperature and pressure, favor flow as opposed to fracture. Jeffreys[43] has expressed it well in his recent book, from which we quote:

> The combined effect of pressure and temperature may be expected to favor continuous deformation, since each has this effect separately; but the stress difference that produces such deformation is raised by pressure and lowered by temperature.

The second part of the quotation is of the utmost importance. It is clear that, with the increased temperatures and the increased

[43] H. Jeffreys, *Earthquakes and Mountains* (London: Methuen & Co., Ltd., 1935), p. 13.

pressures at great depths, the tendency to fail by rock flow rather than by rock fracture is favored. But we have already demonstrated that the increasing pressure-temperature ratio requires that the strength increases with depth, which means that the size of the stress necessary to cause failure increases with depth. We can therefore come to the conclusion that the increased temperatures and pressures at depth may influence the type of yield, but the size of stress necessary to cause distortion of any sort steadily increases downward.[44] At a given depth the distortion may be by flow rather than by fracture, but the strength of the material for the given depth must be exceeded before either type of failure will result.

Now we must answer the question as to why rock fracture occurs rather than rock flow, or whether each may occur. This brings us to the consideration of still another factor which is influential in determining the type of yield. Let us suppose that in a given environment a stress exceeds the strength of the material. We will suppose that the environment is such that the high temperatures and pressures favor a continuous distortion and that rock flow by recrystallization begins. If this rock flow is able to proceed rapidly enough to accommodate the stress or, in other words, if the stress does not build up more rapidly than the material can flow, there will be no fracture and no earthquake. If, on the other hand, the stress is built up more rapidly than the material can reorganize itself by recrystallization, fracturing must result.

There may be a constant rock flowage going on in parts of the earth, perhaps underneath the great seismic belts, but the stress which is causing this flow must, at least at present, be acting a little more rapidly than the material can flow and must be relieved periodically by fracturing.

We shall not attempt to discuss all the significances of these conclusions upon how the material of the earth yields to differential stresses, but there are a few phenomena to which the conclusions apply readily.

The frequency of fractures near the surface of the earth, for ex-

[44] It is interesting to note that Jeffreys, in spite of his recognition of these facts, is instinctively led to a conclusion regarding strength which, if not entirely incompatible to the facts, is certainly not a logical consequence of them.

ample, may be ascribed not only to the relative weakness of the material, but perhaps also to the fact that the material is in an environment where, once the strength is exceeded by the stress, the fracturing type of yield is favored over the flow type. In addition it could be reasoned that the relative infrequency of earthquakes deep in the earth is due not alone to the great strength there, but perhaps also to the partial dissipation of the stress by rock flow.

Where, as in relatively shallow portions of the crust, the temperature has been abnormally raised by magma intrusions, the strength may be locally lowered; in addition the favored type of yield may be changed so that rock flowage occurs where normally rock fracture would be expected, and the size of stress necessary to cause the distortion may be less than for equivalent depths elsewhere. This may explain much of the rock flowage apparent in the root zones of mountain ranges where the strength of the material as well as its type of yield have been affected by nearby intrusions.

It is important to understand that such a weakening of the material cannot be brought about by simple burial, for burial alone increases the pressure as well as the temperature and consequently increases the strength of the material and its ability to resist any kind of distortion. Only if the temperature-pressure balance can be rendered locally abnormal, either through an abnormal rise in temperature or possibly even through an abnormal decrease in pressure, can rock flow be accomplished by smaller stresses than would be required elsewhere at similar depths. Without these abnormal conditions we must consider that rock flow must require greater stresses than are necessary for distortion of strong materials at the surface.

The time-honored expression, "the zone of flow," means, therefore, a zone in which, first, the strength of the material is exceeded by the stress; second, the temperature and the pressure are high enough (for a given material) to facilitate continuous distortion at the expense of fracture; and, third, assuming the first two conditions to be fulfilled, the stress must not be applied too fast to permit the material to undergo rock flow. The concept of an asthenosphere, a truly weak substratum, in which flow can be caused by extremely slight stress-differences, no longer appears valid.

THE BURIED TOPOGRAPHY OF WESTERN OHIO

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ABSTRACT

Examination of thousands of records from wells in western Ohio has made it possible to ascertain the topography of the bedrock surface beneath the glacial drift. Many preglacial or interglacial valleys, hitherto unknown, are now recognized. The most important are the preglacial or interglacial Maumee, Sandusky, Huron, and Teays. According to Tight, the Cincinnati River was a tributary of the old Teays in preglacial time and drained the Miami basin of southwestern Ohio. However, the slope of the bedrock in this area indicates that the drainage in interglacial time, and possibly in preglacial time, was to the southwest, as at present. The course of the Teays (in Ohio) was in a northwesterly direction from Pickaway County, through Madison, Clark, Champaign, Shelby, Auglaize, and Mercer counties into Indiana. The course of the Teays northward through Delaware and Marion counties was impossible, as the bedrock altitude is too high for it to have flowed in that direction. The old surface beneath the glacial drift in northwestern Ohio is remarkably even and not so rough as might be expected. The Worthington (Lexington) and Parker erosion surfaces, well developed in eastern and southeastern Ohio, are represented on the buried lowlands of western Ohio. The preglacial valleys were deepended during the Deep (interglacial) stage preceding the Illinoian, after much disturbance of the drainage by earlier glaciations.

SCOPE AND ACKNOWLEDGMENTS

The area under discussion comprises the glaciated portion of western Ohio. The purpose of this investigation is to discover the nature of the surface which underlies the glacial drift, to ascertain the preglacial and interglacial drainage lines, and to interpret the physiographic history of the old surface.

During the year 1934, geologists and engineers, working on a government project, collected 148,000 well records for the Ohio Water Conservation Commission. Roughly estimated, about one-half of them are from western Ohio. Unfortunately, the project was abandoned when approximately one-half of the state was covered. These records show the well-head altitudes, the depth to the ground water surface, and the thickness of the glacial drift where possible to obtain it. From these data it is possible to calculate the altitude of the old surface. Approximately one-fourth of the records indicate the depth to bedrock. In some cases the bedrock elevations are so nearly alike and so closely spaced that it was found expedient to plot

only the more significant ones, and in a few instances to strike an average.

Approximately 10,000 oil-well records were obtained from the Ohio State Geological Survey, to which they were furnished by the Ohio Oil Company of Findlay, Ohio. The great majority of these come from Wood and Hancock counties; the remainder from Sandusky, Auglaize, Mercer, Wyandot, Lucas, Putnam, and Hardin counties. About 20,000 well records were examined, and 6,000 of the more significant ones were plotted on base maps of western Ohio.

The writer acknowledges the valuable aid given by the Ohio Geological Survey and the Water Conservation Commission, especially by Wilbur Stout and David Warner of these organizations, who made available the well records and gave important information.

PREPARATION OF THE MAPS

In plotting the records western Ohio was divided into four sections, a base map representing each division. With the exception of the larger streams, all detail was omitted. From the government topographic maps, on which were plotted the depths to bedrock, the elevation of the buried surface was obtained. In the majority of cases the well-head elevations are indicated on the records, and the bedrock altitudes were secured by subtracting the thickness of glacial drift from the figure representing the surface elevation. Contour lines, with 100-foot intervals, were drawn through the bedrock elevations to indicate the major physiographic features. Not all the contours were drawn, however, because of lack of data. In the northern half of the area all the contours are shown; but in the Cincinnati area, where the deep, buried channels are as low as 360 feet above tide, the lowest contour drawn is 600 feet above sea-level. Enough contours are drawn to show the general slope of the buried surface and the direction of drainage. The dashed contour lines are conjectural, indicating the probable altitudes.

THE BURIED SURFACE

By use of 100-foot contours the major features of the old surface can be determined. The axis of an old valley, which the writer designates as the preglacial Maumee, extends through Lucas, Ful-

ton, Williams, and Defiance counties. The present Maumee River lies in this drift-filled valley, but its present axis does not everywhere coincide with the one beneath the glacial drift. The old Maumee flowed in an easterly direction through the depression now occupied by Maumee Bay, into a larger stream which doubtless extended eastward through the Lake Erie lowland.

In Williams, northern Fulton, and Lucas counties the slope of the buried surface is in a southeasterly direction toward the axis of the old Maumee valley. In Henry, Paulding, Putnam, Van Wert, and Hancock counties the general slope of the old surface is in a north to northeasterly direction. The broad, north slope of the preglacial Maumee extends into southern Michigan. The highest altitude (940 feet), as indicated by a well record, occurs in the northwestern corner of Williams County. The lowest elevation (465 feet) is located on the old surface near Maumee Bay. The axis of the buried Maumee valley rises from an altitude of 465 feet in Lucas County to 690 feet in Defiance County and to higher elevations in eastern Indiana. This is a fall of 225 feet in a distance of approximately 75 miles, or an average of 3 feet per mile. The average fall of the present Maumee River from Defiance to Maumee Bay is only 1.8 feet per mile.

The preglacial valley slopes up very gradually to a broad upland to the south. One might interpret this old valley as developed in a region worn down to an advanced stage in the cycle of erosion. The difference in elevation between the highest and lowest points on the old surface, or the maximum relief, in Williams, Fulton, Defiance, Henry, Paulding, Putnam, Van Wert, Lucas, Wood, and Hancock counties is 373, 122, 175, 103, 81, 123, 88, 210, 215, and 237 feet, respectively.

The Erie lobe of the great Pleistocene ice sheets passed over the westward extension of the Lake Erie depression at least twice and accomplished some erosion, perhaps enough to make northwestern Ohio somewhat smoother than it would be if only normal erosion by running water had taken place. No deep, narrow channels, such as were developed during the Deep stage in the valleys of the unglaciated area of southeastern Ohio and in the north-central glaci-

ated portion of the state,[1] occur in the preglacial Maumee valley. Perhaps glaciation scoured out the valley, removing or rounding out whatever channel may have been there in the Deep stage. The valleys developed during this stage in western Ohio were doubtless broader and not so deep as those cut during the same time in the plateau section of the eastern part of the state, because the former were cut in weaker rock, nearer the mouths of the master streams in a region of lower elevation.

The rise of the buried surface, from the axis of the old Maumee valley in Fulton County to the southern boundary of Putnam County, approximately 50 miles directly south, is about 180 feet. From central Defiance County to the southern boundary of Van Wert County, a distance of about 40 miles, the rise of the old surface is about 175 feet. These figures indicate that the buried surface has a low average angle of slope.

Lying mainly within Ottawa, Sandusky, Seneca, Wyandot, and western Marion counties is the broad valley of the preglacial or interglacial Sandusky River, apparently a stream of considerable size. This river flowed through the depression now known as Sandusky Bay and was a tributary of the large stream which occupied the Lake Erie lowland in Tertiary time. The rise of the land along the axis of the buried Sandusky valley, from Sandusky Bay to the southern boundary of Seneca County, is approximately 240 feet; and to the divide in Delaware County, about 400 feet, an average gradient of a little more than 6 feet per mile.

The buried surface, from Ottawa County near Sandusky Bay to the northern boundary of Hardin County, rises approximately 400 feet, about 7 feet per mile. The broad divide between the preglacial or interglacial Maumee and Sandusky rivers stands below 600 feet in Wood and Ottawa counties and rises to an altitude of between 900 and 1,000 feet to the south. The gently rising valley slopes, the broad, low divides, and the wide valleys, all suggest a surface such as one might expect in the later stages of the cycle of erosion.

To the east of the old Sandusky River in Erie County lies the buried valley of the preglacial or interglacial Huron River. The

[1] Karl Ver Steeg, "The Buried Topography of North-Central Ohio and Its Origin," *Jour Geol.*, Vol. XLII (1934), pp. 602–20.

axis of this valley extends southward through Huron, Crawford, and eastern Marion counties. East of this valley, in Richland and Morrow counties, lies the buried slope of the Appalachian Plateau. To the north, in western Erie and Huron counties, the old divide is low, but gradually rises to the Appalachian Plateau in southeastern Huron County. The boundary between the buried lowlands of western Ohio and the Appalachian Plateau can easily be traced in a north-south direction through western Fairfield, eastern Delaware, Franklin, and central Morrow counties, curving to the northeast in eastern Crawford, southern Huron, and northern Richland counties. In Ross and Highland counties the margin of the plateau continues in a generally southwesterly direction. The old Huron valley has a rise along its axis of about 400 feet, which corresponds with that of the old Sandusky, the gradient in feet per mile being about the same.

East of the old Huron valley lies another buried depression, now occupied by the Vermilion River. This stream, not shown on the map accompanying this paper, carried the waters from a large drainage area in north-central Ohio. The drainage from Wayne, northern Holmes, and Ashland counties found an exit into the Lake Erie lowland by way of the broad valley at Jeromesville, Ohio.[2] A large stream once occupied this valley, as indicated by its great depth and width, records of oil wells showing a glacial fill of from 250 to 300 feet. The slope of the old valley floor is to the north by way of the buried Vermilion River depression. Unquestionably, a large stream, comparable to the preglacial or interglacial Sandusky and Huron rivers, drained this portion of Ohio. The Black River and Rocky River appear to have had preglacial and interglacial ancestors, as streams now flowing southward in southern Medina County occupy valleys which bedrock elevations, derived from well records, show once drained northward into the tributaries of the above-mentioned rivers.[3] The preglacial and interglacial Cuyahoga

[2] Ver Steeg, "Drainage Changes in the Vicinity of Loudonville, Ohio," *Ohio Jour. Sci.*, Vol. XXXI (1931), pp. 368–76.

[3] Ver Steeg, "The Buried Topography of North-Central Ohio and Its Origin," *loc. cit.*, p. 616.

also flowed into the Lake Erie depression.[4] In eastern Ohio other streams, such as the one which occupied the buried valley of the Grand River and the old Allegheny River, drained northward in Tertiary and interglacial time. The river which occupied the Lake Erie lowland must have been a large one, as it received the drainage of all northern Ohio. The deep, comparatively narrow, buried gorge of the old Cuyahoga at Cleveland and near the town of Independence, 586 feet below the present level of Lake Erie, and slightly below sea-level (13 feet), indicates that the Lake Erie depression in preglacial, and especially in interglacial time, must have been very deep, nearly 600 feet below the present level of Lake Erie. Since the maximum depth of Lake Erie is about 250 feet, glacial deposits of 350 feet, more or less, fill the old valley. The land doubtless stood higher before the advent of the ice age but was depressed by the weight of the glaciers, which may account for the altitude (below sea-level) of the buried valley of the old Cuyahoga. How much the Lake Erie depression was deepened by the glaciers is impossible to determine, and it may be that the bedrock surface is lower than the above-mentioned figure.

In preglacial time western Ohio was made up of the following drainage areas: (1) The old Maumee, Sandusky, and Huron basins, all of which, including others to the east, such as the Vermilion, Rocky River, and Cuyahoga buried valleys, were parts of a larger drainage area which included the Lake Erie lowland, occupied by a large trunk stream which flowed eastward. (2) The area drained by the Teays River. (3) The region of southwestern Ohio now drained by the Miami and Little Miami rivers. This area may have been a part of the Teays drainage basin, the preglacial Cincinnati River draining it and flowing northward, as postulated by Tight. The presence of preglacial or interglacial major divides indicates that Ohio was divided into four large drainage basins.[5] These are the Lake Erie, the southeastern (Teays), the eastern Ohio (old Ohio), and the central (old Licking) drainage basins. The preglacial

[4] Ver Steeg, "Drainage Changes in the Vicinity of Loudonville, Ohio," *loc. cit.*, pp. 614–20.

[5] Ver Steeg, "Erosion Surfaces of Eastern Ohio," *Pan-Amer. Geol.*, Vol. LV (1931), p. 96.

Sandusky, Huron, Vermilion, Cuyahoga, and other rivers which drained into the Lake Erie depression flowed northward from a major divide which can be traced from a point near Barnesville, Ohio, to the divide between Wills and Stillwater creeks. Trending to the northwest, the divide crosses the Tuscarawas River between Port Washington and Gnadenhutten, whence it continues in a northwesterly direction, extending through Holmes, southern Ashland, and Richland counties, disappearing beneath the glacial drift in western Ohio and continuing through Morrow, southwestern Marion, northwestern Delaware, and Union counties. It connects, or is continuous, with the buried upland in Logan County, where the highest elevation in Ohio occurs. Here the divide meets the broad buried valley of the Teays River.

In Clark, Miami, Darke, and Preble counties can be traced the buried divide which separated the preglacial or interglacial Miami and Little Miami valleys from the lowland of the old Teays to the north. The Appalachian Plateau served as the divide to the east of the old Miami basin. The broad divide which separated the old Maumee from the old Sandusky rivers extended from Ottawa and Sandusky counties, through Seneca, Wyandot, and Hardin counties, where it connects and is continuous with the high area in Logan County. The buried upland in Logan and adjacent counties appears to have a central location, at the meeting place of several divides, the streams flowing away from it in all directions except one. The survival of this higher remnant of limestone, which might be considered a monadnock, is explained by the fact that it was located in Tertiary time at the headwaters of the major streams on the divide which separated the drainage basins in western Ohio. Another divide separated the old Maumee basin from the Teays drainage, extending from the divide in Hardin County in a westerly direction through Allen, northern Mercer, and southern Van Wert counties into Indiana.

THE TEAYS RIVER SYSTEM

One of the first workers to study the preglacial drainage of Ohio was W. G. Tight. In his paper[6] he shows the Teays River extending

[6] W. G. Tight, "Southeastern Ohio and Adjacent Parts of West Virginia and Kentucky," *U.S. Geol. Surv. Prof. Paper 13* (1903), Pl. I.

far to the southeast, in the northwestern corner of North Carolina, its drainage basin including southeastern Ohio, a large portion of West Virginia, and part of Kentucky. By a dotted line he indicates the postulated course of the old Teays through western Ohio. More recently Stout and Schaaf[7] reached the conclusion that the Teays River extended eastward over the Piedmont Plateau, perhaps as far as the Fall Line. During Tertiary time it was one of the master streams of the Appalachians, rising in the hills of western North Carolina, flowing through the valleys now occupied by the New and Kanawha rivers, through Virginia and West Virginia, to a point not far from St. Albans, where it turned westward to flow through the abandoned lowland known as the Teays valley. It entered the present Ohio valley near Huntington, flowing to Wheelersburg, whence it continued northward through a valley now occupied by small streams, to a point near Waverly, where it entered the present Scioto valley. Near Chillicothe the valley disappears beneath the glacial drift. According to Tight,[8] there were two large tributaries to the Teays River: the Portsmouth River, which flowed into it at Waverly; and the Marietta River, which joined it south of Waverly, near the Jackson-Pike County line. Tight shows another large tributary, which he called the Cincinnati River, occupying the present Miami drainage basin and flowing in a northeasterly direction past Hamilton and Dayton to join the Teays.

The present writer, however, believes that the course of the Teays was northward in Ross County, curving to the west in Pickaway County and flowing in a northwesterly direction through southern Madison County. At the London Prison farm, a well was drilled 525 feet into glacial material and the bedrock was not reached. The elevation of the bedrock in this buried valley is therefore less than 538 feet above sea-level. This altitude is well below the elevation of the bedrock in the Teays valley at Chillicothe and Waverly, where the floor is exposed. In Pickaway or southern Franklin counties a large tributary joined the Teays at an altitude below that in the old valley at Waverly—in fact, elevations as low as 580–620 feet

[7] Wilbur Stout and Downe Schaaf, "Minford Silts of Southern Ohio," *Geol. Soc. Amer. Bull. 42* (1931), pp. 663–72.

[8] *Op. cit.*

occur in the counties mentioned above. This stream, known as the preglacial Newark River, flowed through the wide, buried valley which passes near Newark, Ohio, and continues through Licking County near Buckeye Lake and through the northwestern corner of Fairfield County. The width and depth of this old valley indicates a large stream.[9] It is well to remember that the great depth of the buried valleys is probably due to interglacial erosion which took place during the Deep stage. These gorges are cut below the Teays valley of the Parker Strath stage.

It is apparent that the Teays River could not have flowed northward through Delaware and Franklin counties. A major buried divide, standing at elevations more than 900 feet above tide, and from 200 to 300 feet above the floor of the Teays valley at Waverly to the south, is strong evidence against the course in that direction. From Madison County, the course of the Teays extended through Clark and Champaign counties, curving to the north through eastern Shelby County, around the western edge of the upland in Logan County. The buried valley bends to the west in Auglaize County at the point where it received the waters of another preglacial or interglacial stream, which the writer calls the Wahpakoneta River. From here the Teays appears to have flowed through the buried lowland, at present occupied by Lake St. Marys and continuing into Indiana. Well records show elevations low enough for the Teays to have flowed in the direction indicated.

In the area northwest of Lake St. Marys, not enough well records are available to give the location of the deep channel. The elevations shown indicate the presence of a broad valley in Mercer County, northwest of Lake St. Marys. Bownocker[10] reported a channel which was traced across Jay and Adams counties in Harrison Township, Blackford County, Indiana. The maximum depth of a drift reported is 430 feet, while outside the channel the drift is much thinner, not more than 50 feet in places. This channel curves to the northwest, where depths of drift of 438 and 440 feet are

[9] R. E. Lamborn, "The Newark Drainage System in Knox, Licking and Northern Fairfield Counties," *Ohio Jour. Sci.*, Vol. XXXII (1932), pp. 449–66.

[10] J. A. Bownocker, "History of the Little Miami River," *Ohio Acad. Sci. Special Papers*, No. 3 (1900), p. 44.

present. The elevation of the bedrock in this part of Indiana would be approximately 400 feet above tide, which is sufficiently low for the Teays to have flowed through the St. Marys depression and the western continuation of the same valley in Indiana. Of course there is the possibility that the deep channel was made in interglacial time, whereas the Teays flowed in a valley of preglacial or Tertiary age (Parker Strath).

Tributaries entered the Teays River at various points, as indicated in Figure 1. Tight[11] shows a large tributary, the Cincinnati River, flowing northward from Dayton and joining the old Teays. Although the writer accepts the possibility of such a stream, there is evidence that the drainage was southward, as it is today in the Miami basin. This evidence is based on the general slope of the bedrock in the Miami basin. Well records do not furnish conclusive evidence of a course to the north, and it appears likely that a divide existed between the preglacial Teays and Miami drainage basins. Nor do they indicate a broad depression which widens northward and is of sufficient size to have accommodated so large a stream. Furthermore, the buried valleys narrow headward; in other words, the pattern of the bedrock formations is such as to suggest that the preglacial drainage was to the south. It is possible that during the Deep stage, after the reversal of drainage from north to south by way of the Ohio, the old Parker valleys were largely destroyed and the present slope of the buried land surface to the south established. For this reason the writer is willing to admit the possibility of the existence of the Cincinnati River as a tributary of the old Teays. If we accept Tight's interpretation that the Miami basin drainage in preglacial time was to the north by way of the Cincinnati River, it might be argued that after the reversal of drainage, as a result of an early glaciation (pre-Illinoian), the present features of the buried topography were developed in the interglacial stage (Deep stage) which followed. The possibility that the Teays River may have flowed into the Miami basin and found an exit to the southwest is not suggested by the bedrock elevations. The buried divide between the Teays and Miami drainage areas and the old valleys in the Miami basin which narrow headward appear to be evidence that

[11] *Op. cit.*

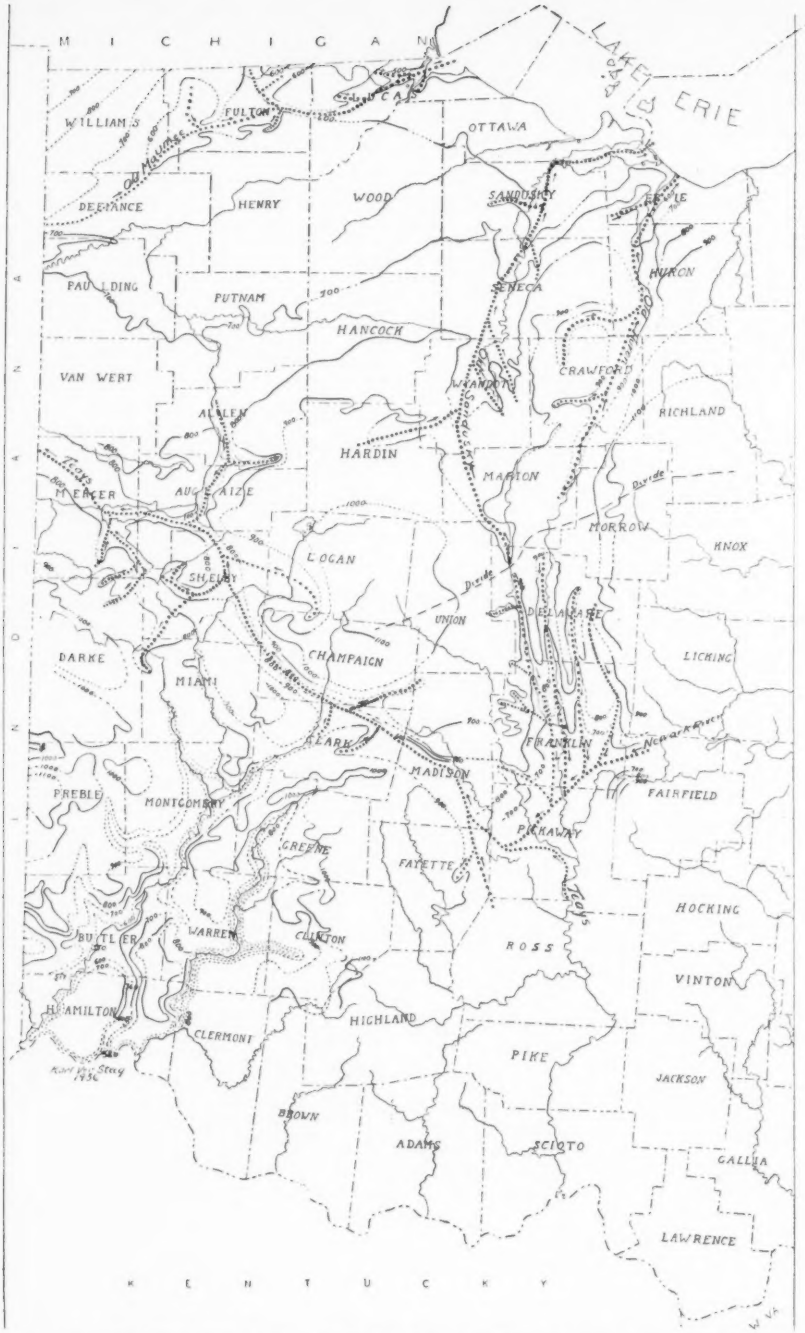

FIG. 1.—Contour map of the buried surface of western Ohio. Contours shown by dashes are conjectural. Large dots indicate the position of the ancient valleys. Contours below 600 feet not shown except in northern portion. Dots are not used to show ancient drainage in the southwestern part.

the Teays River did not have that course in preglacial time. Though admitting that there are arguments for the existence of the Cincinnati River flowing northward as postulated by Tight, the writer points out the alternative interpretation here given.

THE BURIED PENEPLAINS

One could not call the buried surface of western Ohio a rough topography; in fact, it is remarkably smooth or gently rolling in many places, especially in the northern part. The glacial drift, in general resting upon a gently rolling surface, doubtless accounts to some extent for the low relief of the glaciated area in portions of western Ohio. The buried surface in the southern part of western Ohio appears to be somewhat rougher than that in the northern portion. The weak rocks of the early and middle Paleozoic were reduced to a lower level than the Appalachian Plateau to the east, which is covered with the more resistant Mississippian and Pennsylvanian sandstones and conglomerates.

What the surface of Ohio was like at the close of the Permian, immediately following the Appalachian revolution, is difficult, if not impossible, to determine. The land probably sloped westward from the high mountains to the east; and the stream courses, in general, were in the same direction. Later, during Jura-Cretaceous time, Ohio was thoroughly peneplaned, as was the whole of the Appalachian area. Later, uplift and warping along the Appalachian axis from time to time rejuvenated the streams in the plateau area. Denudation probably removed the Devonian, Mississippian, and Pennsylvanian formations from western Ohio, it being possible that these formations once covered a large part, if not all, of that area. The courses of the streams may have been originally determined by the slope of the Appalachian Plateau; and as subsequent erosion removed the upper formations from western Ohio, the western edge of the Mississippian escarpment was shifted eastward and southeastward in the direction of dip. The streams, cutting down into the weaker limestones and shales, developed a broad lowland in western Ohio in Tertiary time. This lowland was continuous with and merged into the great lowland now occupied by Lake Erie and the region known as the Lake Plain immediately to the south. A large

east-flowing stream in the Lake Erie lowland received the waters from tributaries flowing northward and eastward from the Ohio area. This broad lowland was likewise opened up on the weaker Paleozoic rocks which lie below the Mississippian formations. The Lake Erie lowland shared the same physiographic history as western Ohio, and both should be considered as a unit. In the Great Lakes region, where the Paleozoic formations overlap upon the Canadian shield, the beveled edges of the more resistant limestones were brought into relief as cuestas, while the weaker formations were reduced to broad lowlands which, in the pleistocene epoch, were modified into the basins of the Great Lakes.

The question arises as to what erosion surfaces lie beneath the drift cover of western Ohio and the Lake Plain. The writer, in an earlier publication,[12] pointed out that the Lexington plain, in the Blue Grass region of Kentucky, has an average altitude of approximately 900–1,000 feet above tide. This Lexington surface is continuous with the buried surface of western Ohio and with the Worthington lowland along the Ohio River, in the southeastern part of the state, in Lawrence, Meigs, Gallia, Jackson, and Athens counties. In the Cincinnati area, not far from the Ohio-Kentucky line, the elevation of the Worthington (Lexington) surface on the buried uplands varies from 800 to 900 feet above tide. This old surface rises gradually to the north, and in western Preble County stands above 1,100 feet. Its average altitude in Hamilton County is about 800 feet, whereas to the north, in Montgomery, Greene, Miami, and Clark counties, the same surface lies between 900 and 1,000 feet. It slopes up to the west from the buried lowland of the Miami River, rising 200 feet or more in Butler County, its altitude being above 1,000 feet in places. In Preble County the rise to the northwest amounts to as much as 200 feet, its elevation being more than 1,100 feet in the west-central portion. The old surface rises to the east from the buried lowland of the Miami basin, where the altitude averages from between 750 and 850 feet in Warren County, to between 1,000 and 1,100 feet in Clinton and Highland counties. In Greene and Clark counties there is a similar rise of the buried surface to the east toward the upland. It is evident from these facts that in preglacial

[12] "Erosion Surfaces of Eastern Ohio," *loc. cit.*, p. 188.

or in interglacial time there was a broad drainage basin, at present drained by the Miami and Little Miami rivers and their tributaries. The present drainage area of the above-named streams only roughly coincides with the one represented by the buried surface. If this old surface slopes in general to the southwest, what are we to conclude as to the direction of the drainage in this area in preglacial time? Was it to the north, as postulated by Tight, or to the southwest, as at present?

The buried Worthington (Lexington) peneplain was far from being a flat surface, varying from 900 to 1,100 feet in altitude and having an average elevation of between 1,000 and 1,100 feet on the higher areas, on the old divides in Darke, Preble, Clinton, and Greene counties, a relief of about 200 feet. This is not too much to expect on a peneplain, which is in reality a rolling surface, rising toward the divides. The relief of the Schooley (Kittatinny) peneplain, as indicated by profiles on Kittatinny Mountain in Pennsylvania, has a relief of at least 300 feet.[13]

In the area lying within the boundaries of Madison, Fayette, Ross, Pickaway, and Franklin counties the old surface slopes toward the Teays, having an elevation of from 725 to 825 feet near the valley to from 900 to 1,000 feet on the divides. The slope of the buried surface in Delaware County is to the south, the lowest elevations on it being in southern Delaware and Pickaway counties, where the old Newark and the Teays joined and the higher altitudes on it occur to the north on the major divide, where the elevations vary from 900 to 1,000 feet. To the south in Ross County, and to the east in Licking and Fairfield counties, the change from the Worthington (Lexington) lowland to the Appalachian Plateau is rather abrupt. The boundary between the western lowland and the Appalachian upland is rather sharply defined, the change from one province to the other taking place through elevations varying from 900 to 1,100 feet. The boundary between the two provinces might be placed at about the 1,000-foot contour.

The writer digresses here to discuss the possibility of the presence of the lower erosion surface known in southeastern Ohio as the

[13] Ver Steeg, "Windgaps and Watergaps of the Northern Appalachians, Their Characteristics and Significance," *Annals N.Y. Acad. Sci.*, Vol. XXXII, pp. 87–220.

Parker Strath. It is probable that the lower portions of the buried surface along the old streams, at altitudes varying from below 700 feet to 800 feet and above, are in reality another surface cut below the level of the Worthington (Lexington) peneplain. Is the buried surface in western Ohio one peneplain, high on the divides and low near the streams, and only one rolling plain, the Worthington (Lexington); or is it two surfaces, the Worthington (Lexington) on the divides at about 900–1,100 feet, averaging about 1,000 feet, and another, the Parker Strath at lower elevations, somewhat below 700 feet up to 800 feet and above? No sharp line of demarcation exists between the two surfaces, nor would one expect that a sharp division could exist, since erosion tends to remove such evidence. A grada-

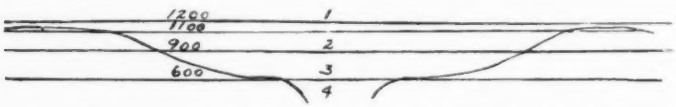

FIG. 2.—Generalized section showing peneplains in western Ohio. (1) Harrisburg surface in Hardin County above 1,100 feet. (2) Worthington (Lexington) surface, 900–1,100 feet. (3) Parker surface, 600–900 feet. (4) Deep-stage valleys cut below the Parker surface. Foregoing figures are approximate.

tion between the two surfaces is what one would expect. Both surfaces are present in the plateau section to the east, the elevation of the Worthington (Lexington) peneplain varying from 800 to 1,100 feet in the driftless area of eastern and southeastern Ohio, an altitude of from 900 to 1,000 feet being the average.[14] The Worthington (Lexington) and Parker Strath surfaces should show slightly lower altitudes in western Ohio since their slope in Ohio is, in general, to the west and northwest. The Parker Strath of western Ohio should be more extensively developed than in the eastern portion of the state because the streams working on weaker rocks, such as the limestones and shales of Silurian and Devonian time, have beveled a much larger area. Furthermore, the vertical distance the streams had to cut down to reach grade or baselevel was less than farther east in the plateau area, where the harder Mississippian and Pennsylvanian sandstones and conglomerates not only stand higher but offer more resistance to erosion. The original slope of the Appa-

[14] Ver Steeg, "Erosion Surfaces of Eastern Ohio," *loc. cit.*, p. 186.

lachian Plateau was no doubt to the west as it is today, and the streams worked headward from the west and northwest. The western part of Ohio was denuded of the harder formations and the weaker rocks exposed at an earlier period. In these less resistant formations the streams have eroded broad surfaces such as the Worthington (Lexington) and the Parker Strath, the latter being so extensively developed that the term "peneplain" might be more appropriate. It is probable that the two surfaces merge on the weak formations so completely that the whole resembles one rolling surface, ranging from below 700 feet to 1,100 feet or more. It would seem unreasonable to assume that the Parker surface does not exist in western Ohio, when not far to the east in the plateau section it is well developed. Surely on the weaker rocks of western Ohio it should be more widespread.

In the eastern part of the state, the Harrisburg surface at 1,200–1,300 feet is preserved on the harder sandstones and conglomerates which constitute the ridges; whereas in western Ohio this surface has been removed entirely, unless we consider that it is represented by the buried upland surface in Logan County, near Bellefontaine. The writer believes that the rolling surface beneath the glacial drift in western Ohio represents two erosion surfaces; the Worthington (Lexington) peneplain on the buried divides, which merges gradually into a lower surface, the Parker peneplain, developed on the buried lowlands along the stream courses, especially the larger ones.

THE DEEP STAGE

Below the Parker surface the streams have incised their valleys, some of which are almost gorges. At the London Prison farm, located northwest of London, in Madison County, Ohio, one well record shows that the drill penetrated 525 feet into glacial material and did not reach bedrock. A very short distance to the east another well record shows a depth of only 324 feet to bedrock. It is evident that here the valley wall is very steep and makes a sharp break with the old surface. In the Cincinnati area well records show similar troughs of great depth, the elevation of the bedrock on the floor of the old channels being as low as 360 feet above tide.[15]

[15] Nevin M. Fenneman, "Geology of Cincinnati and Vicinity," *Geol. Surv. Ohio Bull. 19* (1916), 4th ser., p. 28.

In Allen and Auglaize counties a deep, narrow valley has been cut below an old surface which has an elevation of approximately 800–850 feet near the trough. In Shelby County near the town of Anna, a deep, narrow gorge is indicated by the well records showing more than 500 feet of glacial drift. To the east and west of this channel the elevation of the bedrock is about 800–900 feet. A deep and comparatively narrow valley is indicated by the well records in Mercer County at Lake St. Marys, the surface bordering the trough standing at 700–800 feet. In the northern portion of western Ohio the deep, narrow channels are not so well defined, that is, the valleys appear to be more open and less gorgelike, the valley walls sloping gradually to their floors. Either the cycle of erosion was in a more advanced stage and the valleys were widened to a greater extent than in the southern section, or the erosion accomplished by the glaciers as they pushed up the long slope to the south may have opened widely the valleys cut in the soft limestones and shales of Silurian and Devonian age. Would glacial erosion have removed the narrow channels and opened the valleys to any great extent? The writer believes that the Worthington (Lexington) surface may have been largely, if not completely removed, by normal stream erosion in the northern part of western Ohio and the Lake Erie lowland, the Parker peneplain constituting the buried surface up to altitudes possibly as high as 800 feet. This surface can be traced continuously from northwestern Ohio beneath the glacial drift south of Lake Erie in the Lake Plains province. Below this surface of late Tertiary time was cut a deep valley, later deepened by ice action, partially filled by glacial drift, and now occupied by Lake Erie. This trench was doubtless deepened considerably during the Deep stage by stream action, and probably widened somewhat, as well as deepened, by glacial erosion during the Pleistocene. The islands of western Lake Erie can be interpreted as high points on divides or as limestone hills which stood on the old surface. The peninsula north of Sandusky Bay is probably, in part at least, the unsubmerged portion of an old divide which separated the preglacial Sandusky River from the lowland to the north.

The question arises as to when the valleys below the Parker level were cut. Were they excavated during preglacial time or later dur-

ing an interglacial epoch? Stout holds that the narrow valleys cut below the Parker Strath in southeastern Ohio were eroded during an interglacial epoch earlier than the Illinoian glaciation. According to Stout[16] the Teays (Parker Strath) cycle of erosion lasted until the close of the Tertiary, and the stream valleys were lowered to about 660 feet above sea-level, the approximate altitude of the Parker Strath or Teays bench. The next event was the Kansan or pre-Kansan glaciation, which modified or changed the Tertiary drainage systems. No evidence of this ice sheet has been found in Ohio; but it is believed that the ice from one of the earlier glaciers dammed the lower course of the Teays River and its tributaries, ponding the waters, producing a flood or slack-water stage. Heavy silting took place, and the Minford silts were laid down in the valleys of the Teays system. The ponding was caused by a barrier to the north, either an earlier ice sheet or a rise of the land. The damming of the Teays River would have required an advance of the ice sheet southward into western Ohio as far as Mercer, Auglaize, and Shelby counties, where the land stood at a somewhat lower level to the west of the upland in Logan County. Here the ice would first produce ponding in the Teays valley. The Cincinnati River, a tributary of the Teays, would likewise be ponded if, as postulated by Tight, such a stream existed and flowed northward into the Teays. As a result of the damming of the lower Teays system, the master streams in southern Ohio were reversed in general direction, flowing westward instead of northward as before the coming of the ice sheet. This is known as the Deep-stage drainage and bears little resemblance to the Teays. It cut across many Teays valleys and shows numerous reversals. The master stream gathered waters north and east of Newark, flowed past Buckeye Lake, Circleville, Chillicothe, Portsmouth, and westward to Cincinnati. At Portsmouth it was joined by the Deep-stage Ohio River, the headwaters of which are not determined. This stream cut down rapidly along the lower reaches, at Portsmouth 200 feet below the old Teays valley. The new drainage reached maturity or late youth in the erosion cycle—at least the region was well dissected throughout. It may be that in some places

[16] Ver Steeg, "The Buried Topography of North-Central Ohio and Its Origin," *loc. cit.*, p. 611.

even the uplands were reduced, so that the buried topography really represents a surface largely developed during interglacial rather than in preglacial time. Was the buried surface in the Cincinnati area, in the Miami basin in southwestern Ohio, developed in preglacial or interglacial time? If the preglacial drainage in this area was to the north, as postulated, why does the buried surface have a general slope to the southwest, toward the Ohio River? Either the drainage in this area never was to the north or the buried surface was developed during interglacial time, possibly the Deep stage. There is the possibility that the preglacial surface of Tertiary time was largely removed by later dissection. The reversal of drainage in southwestern Ohio, as a result of the damming and ponding of the Teays and its tributaries, would produce a much shorter route to the sea. Rapid headward erosion would result, deeply dissecting the old Tertiary surface, the Cincinnati area and southwestern Ohio being more dissected and farther along in the cycle than sections at greater distances to the east. Deep valleys were cut below the Tertiary level throughout Ohio, as indicated by the gorges in the Teays or Parker Strath in the unglaciated area and the deep, buried trenches in the glaciated area of western and northern Ohio. The buried gorges of north-central Ohio, as indicated by well records, show that the preglacial topography of the glaciated portion of the Appalachian Plateau in this area, was as rugged as the unglaciated region immediately to the south.[17] The depth of the buried valleys may indicate higher land, possibly rising during the Deep stage, and subsidence with the advent of the later ice sheet. The question arises as to whether the deep dissection during the Deep stage was the result of a rise of the land or a shorter route to the sea by the development of a new outlet by way of the Ohio River to the Mississippi (Tertiary embayment). The return of the earth's crust to isostatic equilibrium during the interglacial epoch probably followed the depression of the land as a result of the weight of the ice sheet. The writer believes that the deep, comparatively narrow valleys buried beneath the glacial drift of western Ohio and incised below an old surface (Parker) were cut during Deep-stage time. These valleys do not appear to be as narrow and gorgelike as those in the plateau area

[17] *Ibid.*, p. 607–20.

because of the nature of the sedimentary rocks, those of the western portion of Ohio being easily eroded shales and limestones, whereas those of the eastern plateau section are sandstones and conglomerates, which offer more resistance to erosion. Just before the advent of the glaciers the cycle of erosion was probably more advanced in western Ohio than in the plateau section, for the following reasons: The drainage was to the west and northwest, and hence the peneplanation was more advanced nearer the mouths of the streams. Furthermore, the initial height of the western portion of the state was probably not as great as that of the eastern part, and the streams did not need to cut down so far to reach grade or baselevel. And the streams were excavating their valleys on weaker rock formations. One would expect the Deep-stage valleys to be more open and less gorgelike in the western portions of the state than in the higher plateau area to the east. The Deep stage lasted from the flood stage of the Teays to the advent of the Illinoian glacier.

The next event was the interruption of the Deep stage by the Illinoian glaciation, with the filling of the valleys by till within the glaciated areas and by outwash beyond the limits of the glacial front. Beyond the margin of the glacier new streams were not developed, and the Deep-stage valleys were undergoing aggradation and were not excavated. Apparently not much work was accomplished in the way of clearing the Illinoian drift and outwash. In places most of the outwash appears to be Wisconsin drift. Hence there may have been more clearance than supposed. The Deep-stage valleys are filled with gravels of the later stage. No Illinoian or Wisconsin drift is present in the Teays valley beyond the margin of the drift. The Scioto valley at Waverly is filled with more than 100 feet of outwash, but it did not reach the level of the Teays valley there. The Wisconsin epoch of glaciation was a repetition of the Illinoian, with the exception of far more outwash and local stream modifications near the border. The last event has been the development of the drainage on the drift, which in Ohio is about 70 per cent unmodified. All that has happened, in the Deep stage, in valleys like the Scioto at Waverly, has been the cutting of the gravel fill to a depth of 30–40 feet, an altitude which marks the upper level of the gravel terrace but which stands well above the Teays level.

Since Illinoian time the Scioto valley at Waverly and south of Chillicothe has been filled to a depth of 140 feet and then eroded to a depth of 40 feet. The amount of erosion accomplished from Teays through Deep-stage time is very large, as the present Ohio valley is cut more than 200 feet below the old Teays valley. The depth of

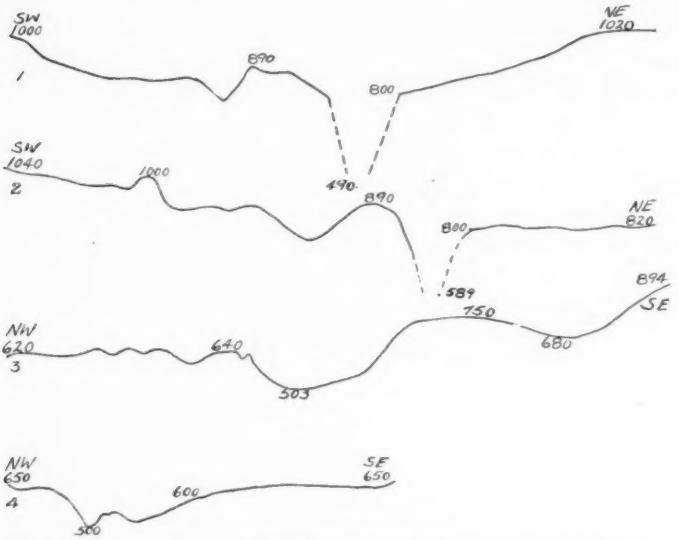

FIG. 3.—(1) Profile across the Teays valley through Darke, Shelby, and Logan counties. Parker surface at 800–900 feet. The Worthington (Lexington) surface at 1,000–1,020 feet. (2) Profile across the Teays valley through Darke and Auglaize counties. Parker surface at 800–820 feet. Worthington (Lexington) surface at 900–1,040 feet. (3) Profile across the Old Sandusky and Old Huron valleys, through Sandusky and Huron counties. Parker surface shown at 600–900 feet. (4) Profile across the Old Maumee valley through Lucas and Wood counties. Parker surface at 600–650 feet.

the gorges and the dissection of the buried surface, especially in the glaciated portion of the plateau in north-central Ohio and in the Cincinnati region of southwestern Ohio, would indicate that a great amount of erosion took place during the Deep stage. As pointed out elsewhere in this paper, it is possible that the present southwesterly slope of the buried surface in the Miami basin of southwestern Ohio is the result of Deep-stage erosion. The Teays drainage represents

only a comparatively small amount of work, as it is simply a deepening of the old valleys on the Worthington (Lexington) surface. Comparing the amount of work done during each period of time, the Deep stage far exceeds the other. But the Teays erosion represents far more work than has been done by the present streams since the ice age. Because of the absence of sufficient data the directions in which the streams which occupied the buried valleys of western Ohio in Deep-stage time flowed, cannot be determined in all cases. The maps indicate the presence of great valleys beneath the drift, but they do not tell us whether they are preglacial or interglacial. It is believed that most of the deep valleys are interglacial in age, that is, they were deepened during the Deep stage.

A summary of the geologic and physiographic history of Ohio since Tertiary time comprises the following events: (1) The development of the Worthington (Lexington) peneplain, remnants of which occur on the buried surface of western Ohio and which can be traced eastward into the plateau section. (2) Uplift and the production of a lower surface known as the Parker Strath in eastern Ohio, but which assumes the proportions of a peneplain in western Ohio, where it represents a more advanced stage in the cycle. The Teays system and the other streams of Ohio flowed in wide depressions representing this stage. (3) The coming of the ice sheet, Kansan or pre-Kansan, and the blocking of the Teays drainage in western Ohio, producing ponding, a flood, or slack-water stage. During this time, heavy silting up to as high as 860 feet occurred. (4) The development of an entirely new system of drainage. The general direction of the master streams in southern Ohio was reversed, flowing westward instead of northward, as before the advent of the ice sheet. This drainage is known as the Deep-stage drainage and bears little resemblance to that which existed in Ohio during the Parker or Teays stage just before the ice age in Tertiary time. (5) The interruption of the Deep stage by the Illinoian glaciation, with the filling-in of the valleys in the glaciated areas by till and beyond the drift borders by outwash. (6) The Wisconsin stage was a repetition of the Illinoian, with the exception of the deposition of far more outwash and of more local stream modifications near the border. (7) The development of the present drainage on the drift.

ON THE STRUCTURE OF THE RHYOLITES IN YELLOWSTONE PARK

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ABSTRACT

Attention is drawn to the relation between steeply dipping lavas and the topography of the surface on which the lavas flowed, also to the relation between topography, drainage lines, and fractures. Much of the banding and lamination of the lavas is considered not as a result of spreading of a non-homogeneous magma but as a result of crystallization along planes developed by movements after the flow structure was completed. Compared with a glacier, the structure of the rhyolite is molded in a less regular way to the form of the surface on which it rests.

The observations are insufficient for a satisfactory discussion of the mode of extrusion of the rhyolites. More information about their structure and its relation to the pre-rhyolite topography may—if anywhere—be obtained from a detailed study of the walls of the deep canyons.

INTRODUCTION

In the summer of 1935 the writer joined the group of geologists which has been carrying on research work in the Yellowstone-Beartooth-Bighorn region during recent years. At Princeton University Geological Research Base under the direction of Professor W. Taylor Thom, headquartering near Red Lodge, Montana, the writer was put into close touch with other geologists working in the region. Yellowstone Park was visited during parts of July and August; the work there was facilitated by Dr. Max Bauer, chief naturalist of the Yellowstone National Park Service, and his staff. The writer was assisted during part of his work by two students of geology, Mr. H. Jansen of the University of Amsterdam and Mr. W. H. Pecora of Princeton University, while Mr. A. Howard of Columbia University, who was finishing his work on the physiography of the Yellowstone River, gave valuable suggestions.

STEEPLY DIPPING LAVA FLOWS

Large portions of the broad volcanic plateau of the central Yellowstone Park consist of horizontal or slightly dipping lava flows, which often show columnar structure. At various places the lavas show lamination and stratification with steep dips. In the region near the falls of the Yellowstone River, in the Firehole Canyon

south of Madison Junction, and in the canyon of the Little Firehole River east of the Upper Geyser basin the general arrangement of steeply dipping lava flows was studied in some detail.

On the banks of the Yellowstone River upstream from the Lower Falls the strike and dip are strongly variable when considered in detail, but there is a more or less regular arrangement of the main strike. It is NW. to NNW. where it crosses the Yellowstone River at the brink of the Lower Falls and turns to WSW. and SSW. farther upstream along the left bank of the river.

In the northern part of the canyon of the Firehole River the main strikes of the steeply dipping lavas are almost parallel to the direction of the river. They turn from NE. to N. and NNW. About 1 km. south of Madison Junction they cross the river and WNW.–ENE. strikes prevail. The lavas often stand on edge or show steep northern and southern dips.

The Little Firehole River flows through a narrow gorge near Mystic Falls west of the Biscuit basin, which is the lower end of the Upper Geyser basin. Vertical and steeply dipping lava flows are clearly exposed at the lower part of the falls, and farther downstream, where the main strike is from NW. to NNW., it locally turns to the NNE. and ENE. Above the falls NNW. and N. strikes prevail, while near the upper end of the gorge the steeply dipping lavas cross the river with a N. and NW. strike.

PRE-RHYOLITE TOPOGRAPHY

In Mount Washburn and Dunraven Peak the early basic breccias which form the base of the rhyolites, rise to great heights above the rhyolites of the central plateau. On their southern margin, in Sulphur Creek, there is an exposure of fine-grained diorite similar to the diorites which occur among the intrusives of Electric Peak and Sylvan Pass. Iddings[1] mentions the rock of Sulphur Creek as the only body of the kind in the central portion of the Park, but farther to the southwest larger and smaller blocks of microquartz-diorite to microgranodiorite, some of which may be *in situ*, cover the slope on the right bank of the Yellowstone River back of the Lower Falls. Between these two localities there is a depression in the pre-rhyolite

[1] J. P. Iddings, "Yellowstone National Park Folio," *U.S. Geol. Surv. Folio* (1896).

topography where the rhyolite of the plateau is exposed with great thickness along the steep walls of the canyon of the Yellowstone River below the Lower Falls. The arrangement of the steeply dipping lava flows, back of the Lower Falls, and the occurrence of older rocks on the right bank of the Yellowstone River suggest that the structure of the rhyolite has been influenced by an eroded surface on which it rests.

That the rhyolite has flowed down in valleys of the pre-rhyolite topography has been mentioned by previous authors. Holmes[2] states that the rhyolites near the mouth of Jasper Creek (now shown on the map as Deep Creek) have, on reaching the brink of the old canyon, poured down its sides and filled in the valley, the line of contact sloping at an angle of nearly 45°. At that period a canyon must have existed where Deep Creek now runs. Along the canyon wall of the Yellowstone River opposite the mouth of Deep Creek there is a tongue of rhyolite extending down the slope, and near the mouth of Agate Creek the pre-rhyolite breccias along the canyon wall are also covered by a tongue of rhyolite, which is inclined with the slope at an angle of 50°.

The slopes of the valleys in the older breccias of Mount Washburn descend steeply to the plateau surface of the rhyolites, suggesting that they had their continuations below this surface before the rhyolite flowed out.[3]

If similar relations prevail where the pre-rhyolite topography is covered by the younger volcanic rocks, the rhyolites flowed on an irregular surface, accumulating in the valleys and depressions, and large portions of the central plateau region may be covered by only a thin crust of rhyolite.

LAMINATION AND STRATIFICATION OF THE RHYOLITES

The rhyolites vary from dark glassy obsidians to reddish or grayish stony rocks with a greater or lesser content of phenocrysts. They have been described in minute detail by Iddings.[4]

[2] W. H. Holmes, "On the Geology of the Yellowstone National Park," *U.S. Geol. and Geogr. Surv. 12th Ann. Rept.*, Part II (1878), pp. 40–41.

[3] *Ibid.*, p. 35; O. T. Jones and R. M. Field, "The Resurrection of the Grand Canyon of the Yellowstone," *Amer. Jour. Sci.*, Vol. XVII (1929), p. 274.

[4] "Geology of the Yellowstone National Park," *U.S. Geol. Surv. Mono. 32*, Part II (1899), chap. x, pp. 356–932.

The lamination in lavas of Yellowstone Park has been explained by Iddings as the result of spreading of the lava during its flow, whereby heterogeneous portions would have become more and more flattened and eventually appear as parallel layers. Philipp[5] has suggested that the lamination in obsidian from Lipari is the result of spontaneous crystallization by increased friction along shearing planes in a highly viscous stage of the cooling mass, before movement had entirely ceased. Much of the banding or lamination in Yellowstone Park is the result of crystallization in a late stage of solidification, when the development of the flow structure was completed and the lava was no longer able to spread. The layers differ in their degree of crystallization, and there is much proof of crystallization *in situ*. Reddish stony rhyolite occurs in layers in the obsidian, but the same rock is also found in isolated bodies of different size without any relation to flow. A large spherical block with a diameter of about 90 cm. occurs in the dark glassy rocks along the main road between the Upper Falls of the Yellowstone River and Chittenden bridge. And in those cases where lithoidal and spherulitic layers in the black glassy rock are so abundant that the black glass forms thin bands or dark lines only, there is no need for the supposition that the lamination was due only to the spreading of a magma which was not homogeneous and contained different amounts of water vapor or gas, but the crystallization may have been greatly favored by circulation of gases along planes developed by movements when the lava had nearly come to rest. Iddings[6] has given proofs of spherulitic crystallization subsequent to cracking of the rhyolite and located along the cracks. He states that spherulitic growth about the edge of an open crack indicates that the magma had been so viscous previous to its last movement that a small gap in it was not closed. It shows that the same mass of rock may be rigid with regard to a force acting through a short period and yet may be plastic toward the same force acting through a longer time. The rate of cooling and the escape of volatile substances will also determine whether lamination and stratification can develop during a shorter or longer period of

[5] H. Philipp, "Beitrag zur Kenntnis der Bewegungsvorgänge in hochviskosen geologischen Flüssigkeiten," *Centralbl. f. Min., Geol., u. Pal.* (1921), p. 679.

[6] J. P. Iddings, "Obsidian Cliff. Yellowstone National Park," *U.S. Geol. Surv. 7th Ann. Rept.* (1885–86), p. 257.

time; these variable factors explain the great variety of banded rock types which are actually found.

The steep dips of the rhyolites in Yellowstone Park often can be easily recognized from a distance by the occurrence of layers of hard, stony, commonly reddish rhyolites in the dark glassy rocks. The layers vary in thickness, which as a rule is not less than several centimeters; they show a more or less complete crystallization of the groundmass, which often has a spherulitic structure.

The late stage in which the spherulites were formed is illustrated in many of the rocks when studied under the microscope. A layer about 30 centimeters thick in the dark obsidian on the left bank of the Yellowstone River, between Upper and Lower Falls to the south of Cascade Creek, shows a flow structure with small overturned and recumbent folds. The fibers of the spherulites were not affected by the flow of the rock nor has the flow structure been affected by the growth of the spherulites, which clearly are of later origin. The spherulites are partly elongated and the fibers then radiate in longitudinal thin section from a line instead of from a central point. These elongated spherulites form a transition to an arrangement in fibrous bands, the fibers being normal to the bands which are more or less parallel to the flow structure and the walls of the layer of hard rhyolite. The spherulites have developed *in situ* and the fibrous bands point to the development of planes of crystallization by shear when the lava was in a highly viscous state. Late movements along planes parallel to the fibrous bands are illustrated by the slickensided surface of sedimentary fragments which occur in the rock.

The orientation of the flow structure and slickensides is given in Figure 1. The plane *ABCD* is parallel to the walls of the layer of stony rhyolite. As is indicated by the minutely folded flow structure, the main direction of flow is parallel to the walls of the layer and to the slickensides, which proves that the movements continued in about the same direction during the cooling of the flow. Thin sections parallel to the planes *ACEF* and *ABGF* show round and elongated spherulites and fibrous bands, the sections parallel to the plane *ABCD* show abundant cross sections of the fibers of the spherulites and fibrous bands, which proves that the spherulites are flattened parallel to the planes of crystallization and to the walls of the layer of stony rhyolite.

FOLDING OF THE RHYOLITES

Besides the minutely folded flow structure which has been mentioned above there is folding of larger dimensions. Resistance in front and below has caused bending in the still-moving mass, where conditions have favored this kind of yielding. The folds are partly overturned. Where the bends of the folds are not exposed, the steep dips of the contact between the dark glassy rocks and the stony rhyolites sometimes suggests the occurrence of dikes. We have studied the sections with this idea in mind but did not find evidence which favors this supposition. On the left bank of the Yellowstone River just back of the Lower Falls the contacts with the stony rocks

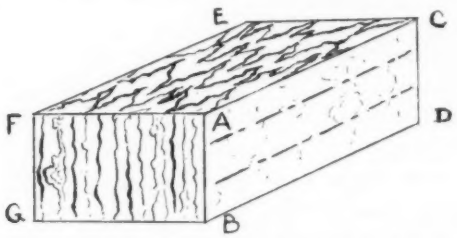

FIG. 1.—Flow structure and slickensides in a layer of a hard stony rhyolite in the obsidian to the south of Cascade Creek. The dashed lines on *ABCD* represent slickensides. Enlargement ×5.

on both sides of the black obsidian dip steeply to the northeast; farther upstream SW. and again NW. dips are found. Near the trail to the rim of the canyon glassy rocks are partly covered by talus, but in places they clearly rest upon the rhyolites. In the higher parts of the slope they form a large exposure between ridges of stony rhyolites. A cross-cutting character of the obsidian could nowhere be proved.

On the left bank of the Yellowstone River downstream from the mouth of Cascade Creek, a steeply dipping to vertical obsidian is strongly eroded away between the walls of harder rocks on both sides; the lamination of the obsidian is partly oblique to the wall.

FRACTURES AND DRAINAGE LINES

Although direct evidence of faulting is scarce, there are features which suggest that movements along fractures are of frequent occur-

rence. The present morphology of the plateau region is influenced by erosion and by the slope on which the rhyolites flowed, but sudden differences in altitude often give the impression that block-faulting and tilting have taken place.

The hot springs are not scattered irregularly; the Lower, Midway, and Upper Geyser basins are arranged from north to south and a more or less regular arrangement is also recognizable for other hot areas.

Fenner[7] remarks that the depression of the geyser basins themselves would—in regions of less abnormal phenomena—suggest the sinking of faulted blocks, but in Yellowstone Park the influence of thermal activity upon rock decay and erosion must be kept in mind. He noted many instances of what seemed to be more than a chance association of hot springs and fumaroles with basins and valleys, where no direct evidence of faulting was perceptible, and he supposes that possibly a combination of several factors has brought about these results.

A glance at the map of the Yellowstone Park shows a fairly regular pattern of the drainage lines, which in the eastern part are more or less from southwest to northeast and from southeast to northwest, while in the western part a tendency to directions from east to west and from north to south prevails. Although fractures are not the only structures which localize erosion channels, the pattern suggests that fractures have influenced the courses of the present rivers.

The fracturing may have been of long duration and a similar pattern may have influenced the drainage lines in pre-rhyolite time. For parts of the Yellowstone River and its tributaries it has been proved that canyons must have existed where the present rivers now run. Pre-rhyolite and post-rhyolite valleys may more or less coincide, but the pre-rhyolite topography may have attained a stage of development with broader and meandering valleys. Steeper dips of the lava flows will be found near the slopes of the pre-rhyolite valleys, while in the main portions of the valleys and above them in the present plateau region the lava may be more or less horizontal. This simple arrangement was complicated by the yielding of the rhyolite

[7] C. N. Fenner, "Bore-Hole Investigations in Yellowstone Park," *Jour. Geol.*, Vol. XLIV (1936), p. 233.

flows to resistances in front and below. Like a glacier with regular cleavage and banding, the rhyolite moves under the force of gravity, but the rhyolite has different properties and is molded in a less regular way to the irregularities of the way. Faulting and erosion during the period of effusion of the rhyolites may also have influenced the simple arrangement.

The main direction of the present Yellowstone canyon may coincide with that of a pre-rhyolite valley, but there will be differences in direction when considered in detail. Supposing a more or less simple arrangement of the dips, the strike of the steeply dipping lavas will cross the present canyon obliquely at those places where its direction is oblique to that of the pre-rhyolite valley (near *B* and *C* in Fig. 2). Different structures of the lavas will then be found in each other neighborhood along the present canyon walls. In the medial portions of the pre-rhyolite valleys the lava may accumulate to great thicknesses, and even in deep cuttings the pre-rhyolite rocks may be visible only where the present river cuts the walls of the pre-rhyolite valley.

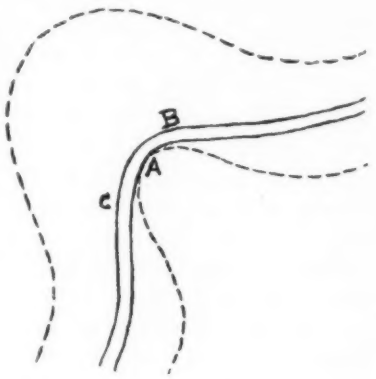

FIG. 2.—Canyon in a pre-rhyolite valley, the wall of which it cuts near *A*.

DIRECTION OF FLOW AND MODE OF EXTRUSION

Sometimes an upward or downward direction of flow movement cannot be inferred from our present knowledge of the steeply dipping rhyolites. We have only very limited information about the structures in the deep cuttings in Yellowstone Park; the known facts do not exclude the possibility that upward movements have taken place along limited sections of the deep canyons. But if the places where the rhyolite broke through the older rocks are visible at the surface, they have not yet been discovered. At many places near the margin

of the plateau the rhyolite occurs in isolated sheets and patches with the structure of surface flows, and the older rocks on which they rest are well exposed. We have not been able to locate any dike, or other intrusive body, in the older rocks when exploring them along the northern and southeastern margin of the rhyolites, and there is no indication that acid rocks which we found among the Sylvan intrusives east of Yellowstone Lake are younger than of pre-rhyolite age.

There are a few places in the Park where the occurrence of dikes and indications of centers of eruption have been recognized, but they seem to be of local importance only. Iddings[8] described dikes of massive-banded lithoidal rhyolite of no considerable extent in the brecciated portion of the rhyolite on the top of Mt. Sheridan in the Red Mountains. This appears to be the only instance in which anything resembling a dike of rhyolite has been observed in Yellowstone Park. Half a mile southeast of Obsidian Cliff[9] he observed a circular pit, the sides standing at an angle of 35°. Its general appearance is that of a small crater; the sides appear to be formed of pumiceous obsidian, the angular masses in the bottom being pumice. Holmes,[10] when descending the western canyon wall of the Yellowstone River near the mouth of Jasper Creek (now shown on the map as Deep Creek), entered a deep gorge which he described as an old volcanic pipe or crater, three sides of which were still preserved, the other having broken down. I have visited the locality but did not find evidence of a center of eruption.

The lack of sufficient evidence makes a discussion of the mode of extrusion of the rhyolites rather unsatisfactory. With regard to the possible modes of extrusion, Daly[11] made the suggestion of extrusion by deroofing of a batholith with a large central area of foundering, the rhyolite passing down, without a break, into granite. It has also been stated[12] that field evidence suggests widespread and

[8] "Geology of the Yellowstone National Park," *loc. cit.*, p. 381.

[9] "Obsidian Cliff. Yellowstone National Park," *loc. cit.*, p. 257.

[10] *Op. cit.* (1878), p. 39.

[11] R. A. Daly, "The Nature of Volcanic Action," *Proc. Amer. Acad. Arts and Sci.*, Vol. XLVII (1911), p. 63. *Igneous Rocks and the Depths of the Earth* (1933), pp. 142–43.

[12] R. M. Field, "Yellowstone-Beartooth-Bighorn Region," *XVI. Intl. Geol. Cong. Guidebook 24. Excursion C 2* (1932), p. 10

progressive fissure eruptions. Where the pre-rhyolite rocks have not been found exposed below the rhyolites we do not know whether such old fissures are present or not. The occurrence of pre-rhyolite rocks on the right bank of the Yellowstone River back of the Lower Falls does not make the area of unknown foundations much smaller.

The desired information about the structure of the rhyolites and its relation to the pre-rhyolite topography may—if anywhere—be expected to be obtainable along the walls of the deep canyons. The very inaccessible character of great parts of these walls has as yet prevented their detailed study. In fact, very little is known about their structure and composition in detail.

REVIEWS

Tectonic Essays Mainly Alpine. By E. B. BAILEY. Oxford: Clarendon Press, 1935. Pp. 200; figs. 49; pls. 5. $4.25.

"The interpretation of the structure of the Alps occupies a prominent position among the advances of Modern Science. The results are amazing, but they are the fruit of long and critical investigation, and have gained general acceptance. I have found it very pleasant to wander up and down the extensive literature of the subject, tracing one eventful discovery after another to its source." These first words of the Preface foreshadow the fascinating stories of Alpine tectonic discovery told with appreciation and charm in this volume. Perhaps in no other field of geologic research can historical narrative be used more advantageously to carry the reader into the present intricate conceptions than in the tectonics of the Alps. In fact, the picture of Alpine structures today is still so unfinished that it is properly appreciated only by seeing the brush-stroking of the principal painters who have outlined and colored it, and the successive erasures and reshapings of various portions as inspiration came from time to time.

Chapter i, "Introduction," and chapter ii, "Chronology of Tectonic Discovery, 1775–1893," sketch the milestones of early progress in the general field of tectonics and prepare the way for a good understanding of the more rapid advances subsequently made in the Alps, particularly the evolution of the nappe theory. For vividness the present tense is extensively employed. Chapters iii, "Switzerland"; iv, "Glarus"; and vi, "Prealps," are pithy descriptions giving essential facts, yet with a light touch which makes easy reading. Chapters v, "Glarus Discovery," and vii, "Prealpine Discovery," resume the narrative of Alpine adventure with critical discussion interwoven. The latter is incisive and illuminating. Numerous little bits of personal equations throw the leaders of the field into sharp relief. The reader comes to know the men as well as their victories and defeats. Three chapters on the "High Limestone Alps of Rhone Sector," "Pennine Alps," and "The Window of the Hohe Tauern" cover selected type areas which have been the scene of some of the most notable structural studies and also display different phases of Alpine development in particularly declared form.

"Marcel Bertrand in Provence" reveals this brilliant geologist in his

workshop. Many glimpses of his rapid succession of papers from 1884 to 1900 show the growth of his tectonic perception and the gradual rise of the nappe conception. The final chapter takes us to Gavernie in the Pyrenees where we witness some sharp debating on the direction of Pyrenean overthrusting during a reunion of French and Swiss masters.

The appreciative reader will be delighted by the skilful word pictures, frequent highlights of both personal and geologic interest, as well as the author's success in making understandable the puzzling geology of the Alps and the bold interpretations which their striking structures have called forth.

R. T. C.

Copper Resources of the World. Vols. I and II. Washington: XVI International Geological Congress, U.S. Geological Survey, 1935. Pp. 885; figs. 140; pls. 41. 2 vols. $10.00; each $6.00.

There are obvious difficulties in summarizing a series of contributions that are in themselves summaries, as are the various chapters of these volumes. All that the reviewer will attempt to do, therefore, is to give some idea of the general nature and scope of the volumes.

The introductory material in Volume I includes a history of the development of the copper industry of the world by J. W. Furness, of the U.S. Bureau of Mines, and the economic history of the copper industry and an estimate of world-reserves of copper by Arthur Notman. This is followed by a chapter on the geologic features of the occurrence of copper in North America by the late F. L. Ransome, in which the principal types of copper occurrence are defined.

The remainder of Volume I is devoted to copper occurrences in North America, beginning with a chapter by F. J. Alcock on copper in Canada. Among the sections dealing with the copper deposits of the United States, those by F. H. Moffit on the copper reserves of Alaska, by Clarence S. Ross on the copper deposits of the eastern United States, the deposits of Arizona by J. B. Tenney, the Michigan copper district by T. M. Broderick and C. D. Hohl, the chapter on the Butte district by L. H. Hart, the chapter on Ely, Nevada, by Alan M. Bateman, and on Bingham, Utah, by J. M. Boutwell are particularly noteworthy.

For Mexico, Manuel Santillán contributes a general chapter, Augustus Locke describes the Boleo copper area of Lower California, and J. B. Tenney describes the famous Pilares mine at Nacozari.

Volume II deals with continents other than North America. For South America, J. T. Singewald, Jr., presents a summary of the outstanding fea-

tures of the famous chalcocite native copper deposit of Corocoro, Bolivia; and Waldemar Lindgren and E. S. Bastin describe briefly the geology of the Braden mine in Chile, which is unique in being a cylindrical ore deposit enveloping an old volcanic vent. The great desert deposit of oxidized copper ore at Chuquicamata, Chile, is described by the company geologist, A. V. Taylor, Jr.; and the large deposit at Potrerillos, Chile, is described by W. S. March, Jr. Cerro de Pasco and Morococha, Peru, are described by D. H. McLaughlin, L. C. Graton, and others. Among the European descriptions, those of Rio Tinto, Huelva, Spain, by Arnold Heim are particularly noteworthy.

In Africa the Katanga deposits are described by Maurice Robert and R. du Trieu de Terdonck. The Northern Rhodesian copper belt is described by Alan M. Bateman. The volume concludes with a chapter on copper-mining methods and costs in North America by E. D. Gardner and C. H. Johnson of the U.S. Bureau of Mines.

In general the volumes are illustrated with many well-selected maps and diagrams and will be an invaluable aid to all geologists concerned with metalliferous deposits, either as teachers or in commercial practice. Space does not permit reference to many shorter sections dealing with deposits that are less well known.

E. S. BASTIN

Invertebrate Paleontology. By W. H. TWENHOFEL and ROBERT R. SHROCK. New York: McGraw-Hill Book Co., 1935. Pp. xvi+511; figs. 175. $5.00.

A book like Twenhofel and Shrock's *Invertebrate Paleontology* has been urgently needed for so many years that it is hard to understand why such a volume has not long since been put on the market. Small wonder, then, that this publication has been met with many enthusiastic comments and instant adoptions. For not only is there no real competing volume, but this work is amply prepared to stand on its own merits even were there the keenest of competition.

The authors have divided their material into ten chapters. The first of these is a well-done, but all too brief, introduction. The succeeding nine deal with twelve invertebrate phyla, the four "worm" groups for convenience being treated together, whereas the bryozoans and the brachiopods are allotted separate chapters. Each phylum is described according to a definite outline, as follows: (1) general features of the subdivisions, (2) biological aspects of the organisms, (3) their hard parts, (4) classifica-

tion, (5) ecology, and (6) stratigraphical range and geological history. Most of the groups are well illustrated with the kind of diagrammatic figures which teachers usually find more valuable than actual pictures of specimens. Moreover, a great many important groups of fossils are discussed which previously have failed to get space in standard texts. Another commendable feature is the book's balance, which must have been obtained only after much careful thought, since even a studied perusal of the volume fails to reveal the authors' "favorite" class of fossils.

It is unfortunate that a publication of such definite merit should be marred by a rather large number of minor errors. Most of these occur in the plates, or their descriptions, and seem to have resulted from the lack of a careful recheck of the drawings after the artist had finished them. A few examples will make this point clear. Figure 50*C* has the blastoid side plate mislabeled; in Figure 56 the *Ipl* of the description turns out to be *Ipi* on the drawing; in Figure 63*A* the ambitus of the exocyclic echinoid has somehow wandered into the trivium; and Figures 116*D–E* are labeled in a fashion calculated only to confuse the student.

Well printed and attractively bound, *Invertebrate Paleontology* to this reviewer does not seem to be overpriced. But, as our British friends chronically complain of any volume that sells for more than twenty shillings, "the cost unfortunately will be somewhat prohibitive to many students."

CAREY CRONEIS

"Zur Geologie der südschwedischen Basalte," by ROLF NORIN, in *Meddelanden fran Lunds Geologisk-Mineralogiska Institution*, No. 57. Lund: Hakan Ohlssons Buckdruckerei, 1934. Pp. 174; figs. 46; map 1.

In the province of Schonen, in southern Sweden, there are about fifty more or less isolated *kuppen* of basaltic rocks cutting gneiss, generally covered at their bases by glacial material, but well exposed above. The individual hills not rarely reach a height of 50 meters, and the basalts usually show columnar structure, many of the individual columns reaching a diameter of $\frac{1}{2}$ meter. The rocks are closely related, and the various types grade into one another. Owing to the lack of sufficient analyses, it was impossible to subdivide these rocks on the basis of the chemical composition; and on account of the presence of glass in so many of the rocks, the mineralogical classification is difficult. A calculation of the mineralogical compositions from the chemical analyses was impossible, since the former cannot always be predicted from the latter. Thus, in the rock from Gella-

berg, which contains nepheline, the content of SiO_2 is 43 per cent, and the normative nepheline 13.6 per cent; while in the rock from Allarpsberg, which contains only feldspar and no nepheline, the normative nepheline is practically the same (13.1 per cent). This led the author to divide the rocks on the basis of their textures into two main groups: those that contain a not unessential amount of glass basis and those that are practically holocrystalline. In the hyaline varieties a brown glass basis with augite microlites dominates; and whatever feldspar there is, is occult in the glass. With further crystallization, the rocks obtain a hyalopilitic texture; and in the case of rocks which have undergone very slow cooling, a hypocrystalline-intersertal texture was developed. The great majority of the rocks, however, are porphyritic, with large phenocrysts in a fine-grained groundmass. Subdivided according to these textures, Norin described petrographically in considerable detail twenty-five different flows. While all of the rocks are called "basalts," actually they include olivine-basalts, nepheline-basanites, and varieties consisting of glass, augite, olivine, and nepheline, which may be basanites or nepheline-basalts, depending upon the composition of the glass.

Many new chemical and modal analyses are given, and there are numerous photomicrographs of thin sections. Descriptions of the various minerals of the basalts are given in separate chapters; and there are other chapters devoted to crystallization and structure, resorption, and the effect of the rocks on magnetic intensities. Observations were made with the intention of determining whether the *kuppen* were independent and distinct bodies—remnants of exposed eruption channels or parts of connected masses. The distribution of the magnetic intensities indicates that the former theory is correct.

A. JOHANNSEN

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